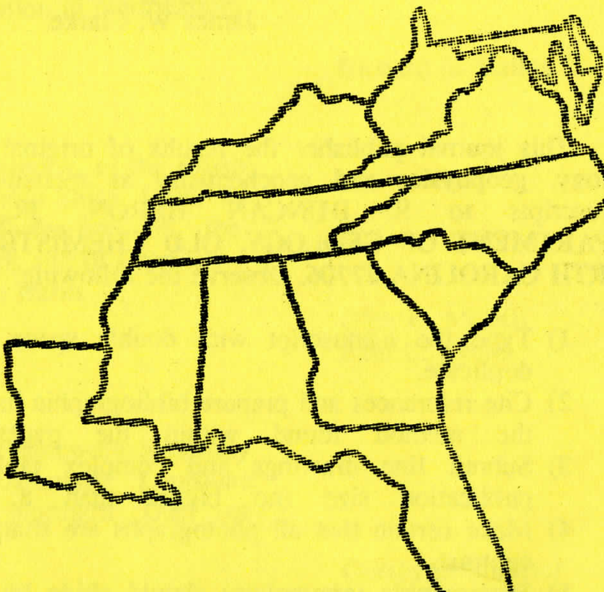
Southeastern Geology: Volume 28, No. 1
August 1987

Edited by: S. Duncan Heron, Jr.

Abstract

Academic journal published quarterly by the Department of Geology, Duke University.

SOUTHEASTERN GEOLOGY



PUBLISHED AT DUKE UNIVERSITY DURHAM, NORTH CAROLINA

VOL. 28, NO. 1 AUGUST, 1987

SOUTHEASTERN GEOLOGY

PUBLISHED QUARTERLY

AT

DUKE UNIVERSITY

This journal publishes the results of original research on all phases of geology, geophysics and geochemistry as related to the Southeast. Send manuscripts to **S. DUNCAN HERON, JR., DUKE UNIVERSITY, DEPARTMENT OF GEOLOGY, OLD CHEMISTRY BUILDING, DURHAM, NORTH CAROLINA 27706**. Observe the following:

1) Type the manuscript with double space lines and submit in duplicate.
2) Cite references and prepare bibliographic lists in accordance with the method found within the pages of this journal.
3) Submit line drawings and complex tables reduced to final publication size (no bigger than 8 x 5 1/8 inches).
4) Make certain that all photographs are sharp, clear, and of good contrast.
5) Stratigraphic terminology should abide by the North American Stratigraphic Code (Am. Assoc. Petroleum Geologists Bulletin, v. 67, p. 841-875).

Subscriptions to *Southeastern Geology* are $11.00 per volume (US and Canada), $13.00 per volume (foreign). Inquires should be sent to: **SOUTHEASTERN GEOLOGY, DUKE UNIVERSITY, DEPARTMENT OF GEOLOGY, OLD CHEMISTRY BUILDING, DURHAM, NORTH CAROLINA 27706**. Make checks payable to: *Southeastern Geology*.

SOUTHEASTERN GEOLOGY

Table of Contents

Vol. 28, No. 1

August, 1987

STRATIGRAPHY AND DEPOSITIONAL HISTORY OF THE YORKTOWN FORMATION IN NORTHEASTERN NORTH CAROLINA

RICHARD H. BAILEY *Department of Geology, Northeastern University, Boston, MA 02115*

ABSTRACT

The outcropping Yorktown Formation in northeastern North Carolina records an asymmetric early Pliocene transgressive and regressive (progradational) cycle. Grain size distributions, bedding characteristics and fossil accumulations, coupled with a chronostratigraphic framework based on mollusks and ostracodes, allow reconstruction of the following depositional history: (1) Initial rapid transgression and accumulation of a reworked sand sheet. (2) Brief regression and/or stillstand. (3) Renewed transgression and deposition of silty shelf sands. (4) Progradation (regression) of a shelf, lagoonal, and possibly deltaic mud blanket; silt and clay bypassed coast marginal environments during a period of slowly falling sea level; termination of progradation by rapid sea level fall.

Both transgressive and regressive deposits show evidence of possible superimposed minor sea level fluctuations, episodic storm surge, and normal wave and tidal current events.

INTRODUCTION

The Yorktown Formation is an areally extensive early to late(?) Pliocene lithostratigraphic unit found in natural exposures from the Rappahannock River in Virginia to the Neuse River in North Carolina. The fauna of the Yorktown, especially the mollusks, has been intensively studied over the past 150 years. Only in the past 15 years or so have the internal biostratigraphic and lithostratigraphic relationships been clarified and applied in a consistent and useful way (Hazel, 1971, 1977; Ward and Blackwelder, 1980; Blackwelder, 1981b). Most of this recent detailed work has focused heavily on the excellent outcrops along the James and York Rivers in southeastern Virginia. The purpose of this paper is to utilize this new framework in a detailed synthesis of the lithostratigraphy, sedimentology, and biostratigraphy of the Yorktown Formation in northeastern North Carolina. This synthesis provides the basis for interpreting paleoenvironments and depositional history during the Yorktown transgressive/regressive cycle.

REGIONAL STRATIGRAPHY

In the area included in this study (Figure 1) the Yorktown strikes approximately N 30 E and dips to the southeast at 0.2 to 0.4 m per km. Thicknesses in most natural exposures range between 2 and 12 m. Toward the west, successively older Coastal Plain strata (upper Miocene to Cretaceous) are onlapped until the Yorktown rests directly on crystalline rocks of the Eastern Slate Belt (Brown, 1985). Locality numbers used in text refer to Figures 1 or 4 and

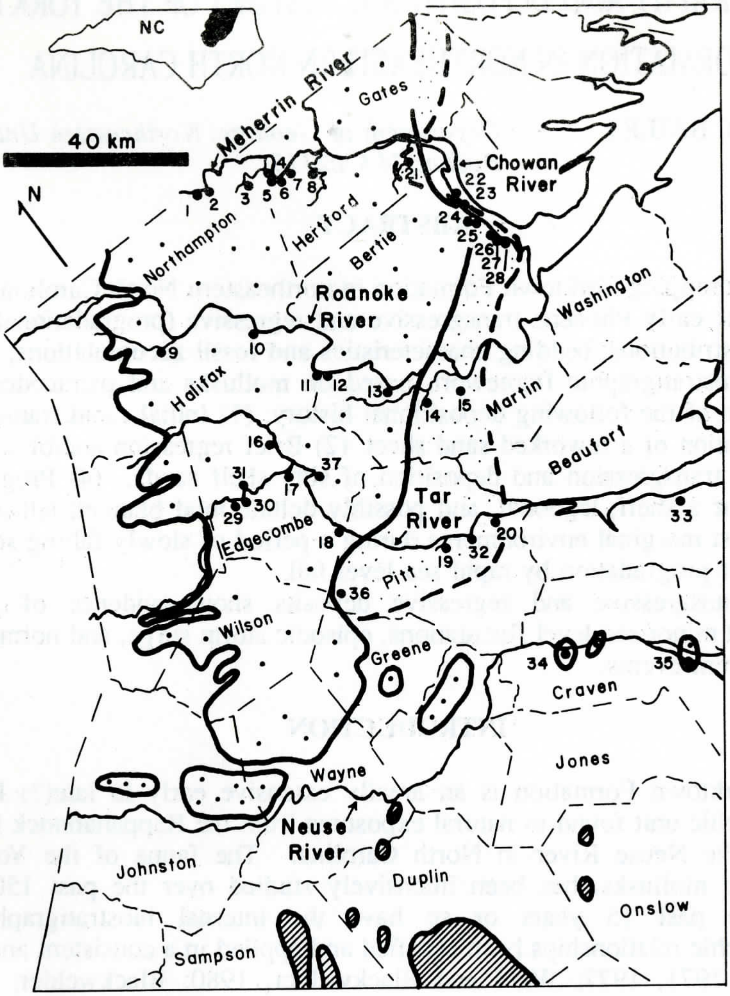


Figure 1. Portion of Coastal Plain of North Carolina showing localities of Pliocene outcrop in northeastern N.C., small inset map shows outcrop area enlarged in lower diagram, Yorktown Formation outcrop area (heavy stippling) enclosed by solid lines, Duplin Formation outcrop area ruled, Chowan River Formation outcrop area lightly stippled and enclosed by dashed line. Detailed locality information is given in Appendix.

and detailed locality information is given in appendix. In Bertie and Hertford Counties (localities 21 and 22), the late Pliocene Chowan River Formation unconformably overlies the Yorktown. A similar relationship may exist in some areas of Martin and Pitt Counties but the contact has not been observed. Throughout the rest of the outcrop area, Yorktown strata are unconformably overlain by Plio(?)- Pleistocene fluvial and estuarine(?) sediments. The upper contact is often marked by a pebbly ferricrete layer. In some areas there was substantial erosion of Yorktown deposits prior to fluvial deposition. At locality 10 on the Roanoke River (Halifax County) all but about 10-15 cm of Yorktown sediments were eroded. The sediment that is preserved occurs in burrows and shallow depressions on an undulatory upper surface of the Cretaceous Cape Fear Formation.

As redefined by Ward and Blackwelder (1980), beds at locality 4 on the Meherrin River (Hertford County), previously assigned to the St. Marys Formation (Mansfield, 1944) and the lower Yorktown (Gibson, 1970), are assigned to the Cobham Bay Member of the Eastover Formation. Beds previously considered to be upper Yorktown (Mansfield, 1944) along the Chowan River (localities 21 - 28), the Roanoke River (localities 14 - 15), the Tar River (19, 20, 32), and near the Martin Marietta Quarry at Fountain, N.C. (locality 36) are now assigned to the Chowan River Formation (Blackwelder, 1981a). The Eastover, Yorktown, and Chowan River Formations comprise the upper part (Miocene and Pliocene) of the Chesapeake Group of Maryland, Virginia, and North Carolina. The distribution of the formations of the Chesapeake Group reveals the presence of regional basins that influenced deposition and recorded the shift of a late Oligocene to middle Miocene depocenter from the Salisbury Embayment in southeastern Maryland and northeastern Virginia to the Albemarle Embayment of southeastern Virginia and northeastern North Carolina in the late Miocene and Pliocene (Mansfield, 1929; Gibson, 1970, 1983). A southeasterly trending structural high in the vicinity of the Neuse River was a positive feature in Neogene pre-Yorktown time that formed a portion of the southern edge of the Albemarle Embayment (Gibson, 1967, 1983). Yorktown sediments are either missing along this structure or they are thin, high energy deposits over Eocene and/or Oligocene strata. At locality 35 near New Bern a 1.8 m bed of highly comminuted shell and pebbly, silty sand of upper middle Yorktown (Rushmere Member) is found overlying Oligocene and Eocene rocks.

To the south of the Neuse River in Jones, Lenoir, Wayne, and Johnston Counties strata of Yorktown age occur as isolated erosional outliers. The name Duplin Formation is applied to the lithic unit in southern North Carolina (generally south of the Neuse River) and northern South Carolina that is biostratigraphically equivalent to the Yorktown. Blackwelder and Ward (1979) suggested that the name Duplin be abandoned and the Yorktown extended to include these coeval strata. I believe that the name Duplin Formation is a valid and useful stratigraphic unit that should be retained for the following reasons: 1. Duplin beds occur south of the Albemarle basin that contains the Chesapeake Group and they are not mappably continuous with Yorktown strata. Ward and Blackwelder (1980, p. 40) state essentially this. 2. Coarse, highly bioclastic-rich Duplin sediments on and south of the Cape Fear Arch were deposited under subtropical conditions and frequently in very shallow marine paleoenvironments. At many, though not at all localities, Duplin strata are lithologically distinct from Yorktown strata. Blackwelder and Ward (1979, p. 35) also describe these characteristics and in addition point out that lithologic subdivisions of the Yorktown cannot be applied to the Duplin. 3. The term Duplin Formation is well entrenched in the literature and, with proper definition, is as valid and useful as any other rock stratigraphic unit of the Chesapeake Group. Abandoning the term Duplin Formation will make discussion of rock units and paleogeography needlessly more intricate.

YORKTOWN LITHOSTRATIGRAPHY

The Yorktown Formation in southeastern Virginia is subdivided into four lithologically distinct members, which in ascending order are the: Sunken Meadow, Rushmere, Morgarts Beach, and Moore House (Ward and Blackwelder, 1980). The Sunken Meadow, Rushmere, and Morgarts Beach Members are well exposed along the Meherrin River and its tributaries. Along the Roanoke and Tar Rivers only the Rushmere and Morgarts Beach Members are found, and their lithologies are less diagnostic than farther to the north. The Moore House Member has not been observed in northeastern North Carolina.

In the following discussion the lithology of each member is summarized, the map distribution within the outcrop area is given, and correlations of members in the Meherrin, Roanoke, and Tar River valleys are evaluated.

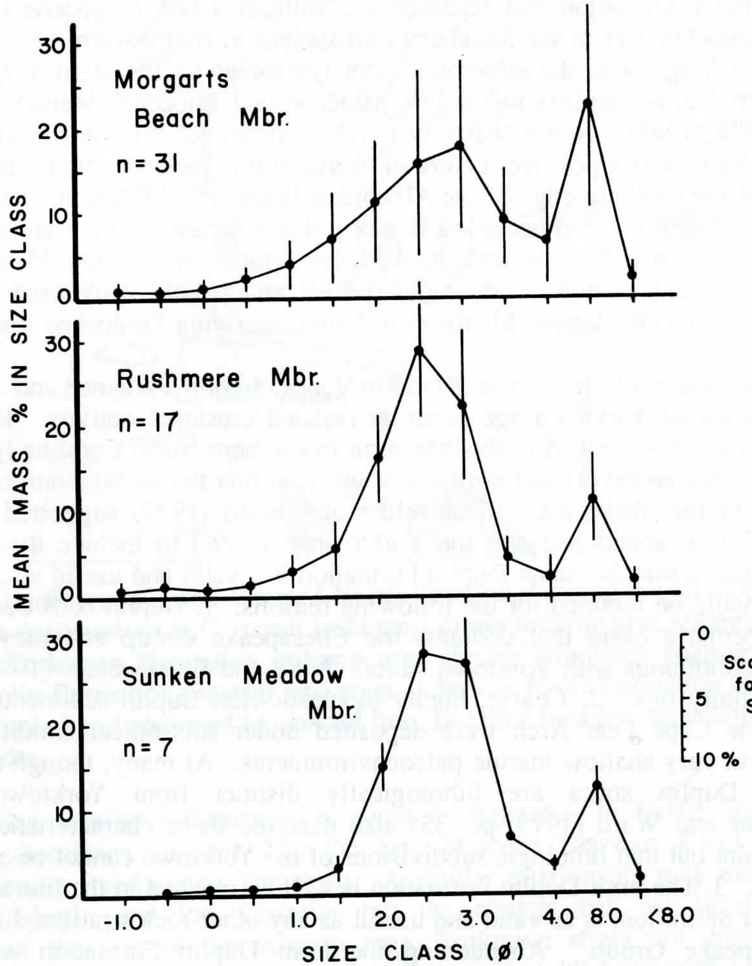


Figure 2. Summary of 55 grain size analyses of Yorktown sediments. Number of analyses summarized in each diagram is given under member name. Vertical bars represent 1 standard deviation, scale for standard deviation on lower right.

Sunken Meadow Member

The Sunken Meadow Member is exposed at localities 3,4,5, and 6 along the Meherrin River and Kirby Creek in the vicinity of Murfreesboro in Hertford and Northampton Counties. This unit pinches out to the south and west where the overlying Rushmere Member lies directly on pre-Yorktown rocks. Downdip at locality 33, Sunken Meadow strata overlie the middle Miocene Pungo River Formation (Gibson, 1983). The maximum exposed thickness at locality 4 is 3 m. At all localities in North Carolina, and in the type area in Virginia, the typical lithology is a fine, moderately to well sorted sand containing about 5 - 15 % silt and 1 % clay. Figures 2 and 3A summarize size analyses of 7 samples from the outcrop area in North Carolina and several samples from the type area in Virginia. Primary sedimentary structures are lacking; however, burrows are common as are thin 5 to 20 cm thick beds of horizontal pectens. Burrows, 2 to 5 cm in diameter, are filled with fine to very fine silty sand and small shell fragments. Articulated bivalves, *Chesapecten*, *Glossus*, "*Dinocardium*", and *Cyclocardia* are common, and especially abundant in the Sunken Meadow Member immediately overlying the Eastover Formation at locality 4.

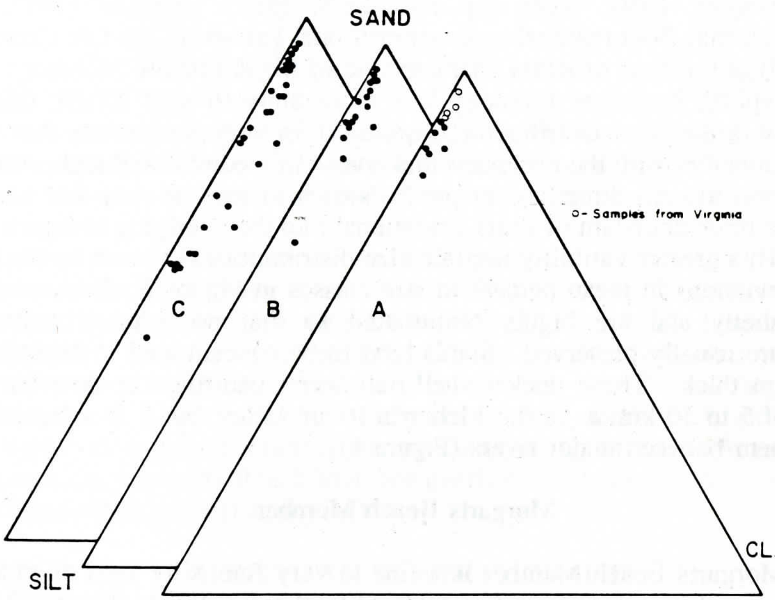


Figure 3. Sand, silt, and clay proportions of sediment samples from Yorktown members. Sunken Meadow, A; Rushmere, B; Morgarts Beach, C.

Rushmere Member

The Rushmere Member extends throughout the Yorktown outcrop area of northeastern North Carolina where it is found in outcrops in the Meherrin, Roanoke, and Tar River valleys. To the south and west it oversteps the underlying Sunken Meadow Member and rests on Paleozoic granite at locality 9 and on Cretaceous strata along the Tar and Roanoke Rivers at localities 10, 11, and 37

(Figure 4). Isolated outliers of the Rushmere also overlie Eocene and Oligocene rocks along the Neuse River (Figure 1). The exposed thickness of Rushmere strata ranges between 0.5 to 2.7 m. In the Meherrin River valley there is a 0.3 to 0.6 m thick shell bed near the base and top of this member. Molluscan fossils, although often concentrated into beds, are usually very well preserved, commonly articulated, and unabraded. In the basal shell bed, larger specimens are often heavily bored and *Chesapecten* valves may be covered with large clumps of the barnacle *Balanus*. At localities 3, 4, and 5 where the Sunken Meadow/Rushmere contact can be studied, it is a subtle transition that appears to be completely gradational. Fossils characteristic of the underlying Sunken Meadow, such as *Placopecten clintonius* (Say, 1824) and *Chesapecten jeffersonius* (Say, 1824) occur in the basal shell bed as do abundant phosphatic grains and small rounded pieces of cetacean bone. These features suggest a period of submarine winnowing and/or non-deposition between the Sunken Meadow and Rushmere Members. Along the James River in Virginia, Sunken Meadow strata are locally and regionally truncated by the overlying Rushmere Member (Ward, L.W.; personal communication). This relationship suggests that there is an unconformity between the Rushmere and the underlying Sunken Meadow in southeastern Virginia (Ward and Blackwelder, 1980; Ward and Strickland, 1985). Gibson (1967, 1983) reported what may be a regional scour or erosional surface in the Lee Creek mine (locality 33) at the base of strata characteristic of the Rushmere Member.

The typical Rushmere lithology is a moderately to well sorted, fine sand. The general grain size distribution (Figure 2) is very similar to that of the underlying member with the exception that coarse to medium sand and pebbles are more common in beds directly over pre-Yorktown rocks, and very fine sand, silt, and clay are more abundant in strata gradational into the overlying Morgarts Beach Member. This greater varibility in grain size distributions is shown by the greater standard deviations in mean percent in size classes in Figure 2. Rushmere sands are very shelly and are highly bioturbated so that no primary sedimentary structures are usually preserved. Shells tend to be concentrated in distinct layers 0.2 to 1.2 m thick. These thicker shell-rich layers can often be correlated over distances of 5 to 10 km, as in the Meherrin River valley, but it is not possible to correlate them between major rivers (Figure 4).

Morgarts Beach Member

The Morgarts Beach Member is a fine to very fine, silty sand or sandy silt. Silt and clay content ranges from 5 to 60 % of total sample mass (Figures 2 and 3). Along the Meherrin and Roanoke Rivers the Morgarts Beach Member consists of a thick sequence (up to 10 m exposed) of thinly to thickly laminated sandy silt. Silt and clay laminae (0.1 to 0.8 cm thick) are separated by very thin (0.1 to 0.3 cm) laminations and lenses of very fine sand. In many areas the laminations have been partly or completely destroyed by bioturbation. The bivalves *Mulinia congesta* (Conrad, 1833), *Yoldia laevis* (Say, 1824), and *Nucula proxima* (Say, 1820) occur in great abundance in these fine sediments, often to the almost total exclusion of other macrofauna. Coral and large bivalves such as *Chesapecten* are occasionally common in this member at some localities. Beds of shell hash 0.3 to 0.9 m thick are composed of abundant disarticulated and comminuted *Yoldia* and *Mulina*

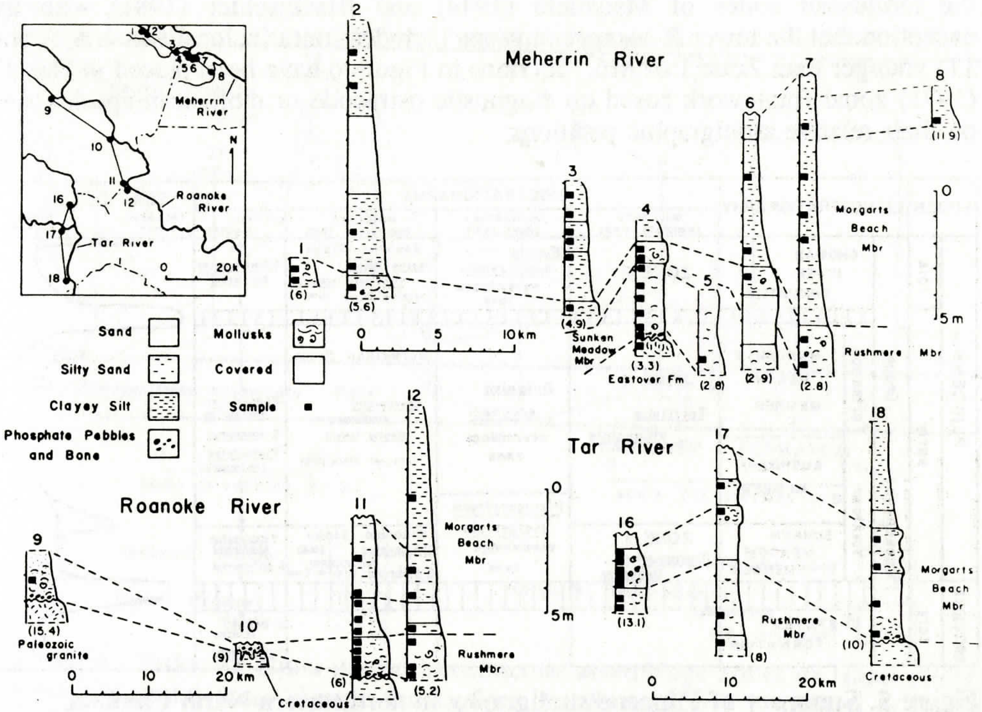


Figure 4. Correlation of Yorktown sections along the Meherrin, Roanoke, and Tar Rivers. Inset map shows lines of sections. Number in parentheses at base of each section is elevation of base of exposed Yorktown in meters.

valves. Convex-up *Carolinapecten* and *Chesapecten* are common in these beds as are large specimens of *Panopea* in living position in shell filled burrows. Small pyrite crystals and diatom frustules are locally abundant in the laminated clayey silt. Along the Tar River in Edgecombe County (localities 16-18) strata assigned to the Morgarts Beach Member consist of beds of very fossiliferous fine sand often interbedded with *Mulinia*-bearing sandy silt. These sandy beds along the Tar River appear to represent a higher energy facies of the Morgarts Beach Member. In that area the Morgarts Beach Member overlies the Rushmere Member (locality 17) or Cretaceous strata (locality 18).

YORKTOWN BIOSTRATIGRAPHY

Mansfield (1944) divided the Yorktown into a lower *Placopecten clintonius*, Zone 1, and an upper *Turritella alticostata* (Conrad, 1834), Zone 2. Blackwelder (1981b) designated Mansfield's lower zone his M6 or *Chesapecten jeffersonius - C. madisonius* (Say, 1824) interval zone and the Wiltonian Stage (Figure 5). Strata of Mansfield's upper zone were named the Burwellian Stage and defined as the M5 or *Chesapecten madisonius - Noetia limula* (Conrad, 1832) interval zone (Blackwelder, 1981b) (Figure 5). Hazel's (1971) lower *Pterygocythereis inexpectata* (Blake, 1929) and upper *Orionina vaughani* (Ulrich and Bassler, 1904) ostracode assemblage zones of the Yorktown are very approximately equivalent to the molluscan zones of Mansfield (1944) and Blackwelder (1981) with the exception that the lower *P. inexpectata* zone includes strata (at localities 4, 6, 9, and 11) younger than Zone 1 or M6. Sections in Figure 6 have been placed in Hazel's (1971) zonal framework based on diagnostic ostracode or molluscan species, and on their relative stratigraphic positions.

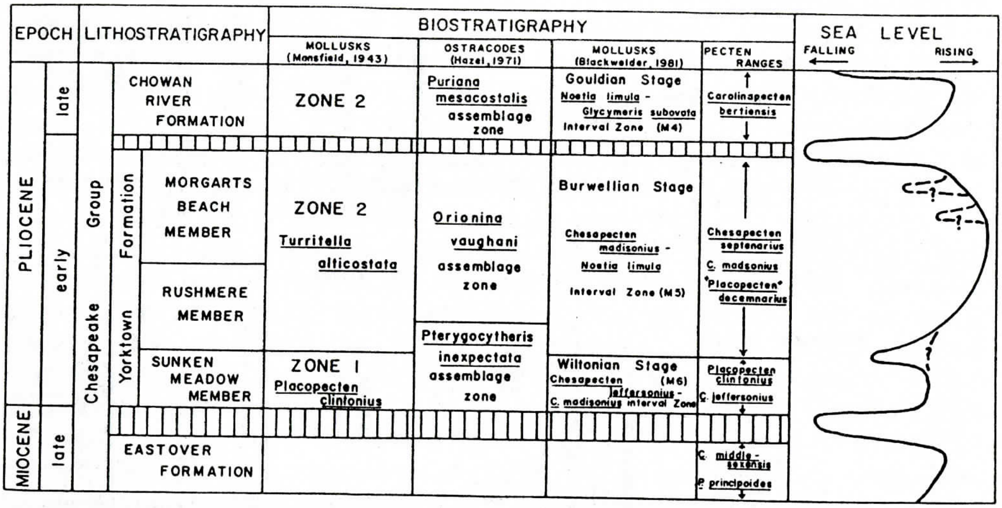


Figure 5. Summary of Pliocene stratigraphy in northeastern North Carolina.

The fact that Hazel's (1971) *Pterygocythereis inexpectata* assemblage zone includes the Sunken Meadow and the lower part of the Rushmere Member (as at localities 4, 6, 9, and 11) requires explanation. In his original definition, the *P. inexpectata* zone was based on 10 samples; 2 or 3 from the Sunken Meadow Member and 7 or 8 from the lower part of the Rushmere Member (Hazel, 1971; p. 3). Thus the *P. inexpectata* zone, as defined by Hazel (1971), may span a significant stratigraphic discontinuity, and is largely characterized by a composite suite of moderate to deep shelf species from the basal Rushmere (= lower *Turritella alticostata* zone) and the Sunken Meadow Members (= *Placopecten clintonius* zone). Hazel's *Pterygocythereis* and *Orionina* assemblage zones appear to be equivalent to Bailey's (1973) *Chama-Crepidula* and *Mulinia-Yoldia* molluscan assemblages of the Rushmere and Morgarts Beach Members respectively. These molluscan assemblages were controlled by environmental differences, primarily depth, temperature, and bottom type, associated with the deposition of the Rushmere and Morgarts Beach Members (Bailey, 1973). The Yorktown assemblage zones seem to be facies dependent. Many of the biostratigraphically useful ostracode species range completely or largely through the Yorktown (Hazel, 1977) and no listed ostracode species is restricted to the Sunken Meadow Member and Mansfield's zone 1. All of these factors render Hazel's concept of chronostratigraphically significant ostracode assemblage zones a difficult one to apply to detailed stratigraphy of the Yorktown Formation. Similar problems probably exist for some of the molluscan taxa used as index fossils for particular stages or zones.

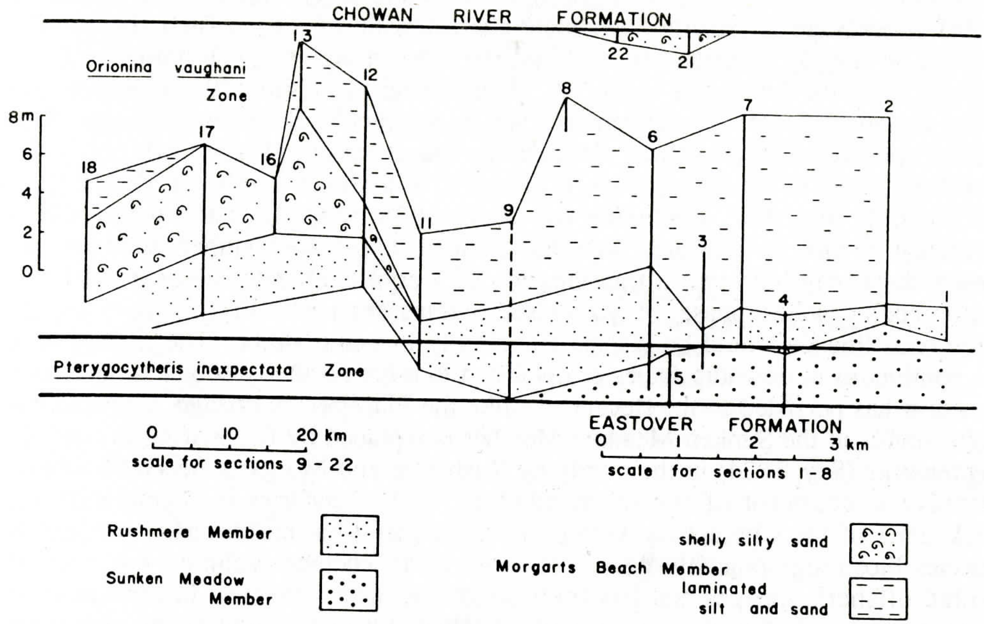


Figure 6. Yorktown strata placed in ostracode assemblage zones of Hazel (1971). Note horizontal scale change for sections 1 - 8; section 7 is plotted in correct position and 1 - 8 are plotted relative to 7 using horizontal scale indicated. All sections are projected into plane trending N 20 E or approximately parallel to strike of Yorktown Formation.

The Zone 1 - Zone 2 or M5 - M6 boundary corresponds to the contact between the Sunken Meadow and Rushmere Members. As discussed earlier, this contact has been interpreted as an unconformity (Ward and Blackwelder, 1980; Ward and Strickland, 1985) caused by a rapid and brief regression. At isolated localities in the Meherrin River valley this is a subtle contact with little physical evidence for a prolonged regression. Similarity of sediment grain size distributions for the two adjacent members (Figure 2) and the lack of diagenetic or physical structures indicating subaerial exposure support the alternative explanation that the M6 (Zone 1) fossil assemblage and strata represent an initial transgressive phase of a longer term transgression-regression cycle. A rapid early transgression followed by a stillstand or slow fall of sea level (or decrease in the rate of sea level rise) could have produced a condensed stratigraphic section over a submarine scour surface (Seilacher, 1982). During the formation of this surface, some M6 species became extinct or migrated out of the preserved basin. There is also evidence that the M6 assemblage represents a mild temperate or deeper water offshore community with certain species, most notably *Placopecten clintonius* and *Chesapecten jeffersonius*, eliminated by migration and pseudo-extinction, respectively, when the marine paleoclimate changed or the basin shoaled during a stillstand. The M6 (Zone 1) assemblage could thus be, at least in part, an ecologically controlled biostratigraphic unit. When the transgression accelerated in mid-Yorktown time the cooler water species existed offshore or at greater northern latitudes. This idea gains some support from two other lines of evidence.

Gibson (1967, 1983) in studies of benthic foraminifers, reported that the Yorktown (M6) assemblage at locality 33 is most characteristic of a deep shelf (depth 80 - 100 m or slightly deeper). In addition, the close morphological similarity of *P. clintonius* to the living sea scallop *P. magellanicus* (Gmelin, 1791) indicates that they are ancestor and descendant species in a direct evolutionary lineage. The extant species inhabits shallow shelf environments north of Cape Cod, but at the southern limits of its range (Cape Hatteras) it is found at depths of 55 to 110 m (Stanley, 1970). If *P. clintonius* was a cryophilic species, as the foraminiferal evidence seems to indicate, then the Sunken Meadow Member and the M6 assemblage may represent a transgression of a cool-temperate sea or possibly an early cooler water phase of an episodic transgression. In this case the *P. clintonius* lineage would be discontinuous only in updip strata. The lineage may be continuous in downdip facies or displaced to more northerly cooler water facies where it has persisted to the Recent. Unlike the *Placopecten* lineage, *Chesapecten jeffersonius* in the Sunken Meadow Member is replaced by *C. madisonius* and *C. septenarius* (Say, 1824) in the overlying Rushmere and Morgarts Beach Members. Iterative replacement of species at lithologic unit boundaries is characteristic of rock units of the Chesapeake Group and is in fact what makes many molluscan species biostratigraphically useful. It is not clear whether evolution was gradual within offshore lineages and has been discontinuously preserved in transgressive sequences, or whether the lineages are punctuated by speciation events that were causally related to environmental factors associated with transgressive-regressive cycles and therefore are coincident with lithic unit boundaries.

DEPOSITIONAL HISTORY

The subcrop and facies maps (Figures 7 and 8) summarize the depositional history and facies distributions of the Yorktown Formation. In early Yorktown time (lower *Pterygocythereis inexpectata* assemblage zone) a sea transgressed to the southwest and west out of a restricted basin or embayment in southern Virginia and northeastern North Carolina. During this initial transgression the Sunken Meadow Member was deposited (Figure 8A). There is a major biostratigraphic discontinuity between the Sunken Meadow and the underlying Cobham Bay Member of the Eastover Formation (Figure 9). I interpret this contact as a disconformity or at the very least a zone of significant stratigraphic condensation (Seilacher, 1982). The latter could occur in part during a prolonged episode of slow deposition and/or submarine sediment winnowing. At locality 4 a bed of densely packed articulated *Isognomon* valves and several species of pectens of the uppermost Eastover is overlain by a 5 to 20 cm layer of silty sand containing abundant articulated bivalves, especially "*Dinocardium*" and *Placopecten principoides* (Emmons, 1858). I place the Yorktown - Eastover contact at the top of the "*Dinocardium*" bed and below a thin lag of sparse reworked *Isognomon*, *P. principoides*, whale bone, phosphate pebbles, and shark teeth. Upward mixing of the Eastover fauna with species typical of the basal Sunken Meadow Member and similar grain sizes of Eastover and basal Yorktown sediments result in a subtle and apparently gradational contact at locality 4. The uppermost *Isognomon* bed of the Eastover Formation in both North Carolina and Virginia (Bick and Coch, 1969) often has a unique edgewise packing of large bivalves (Figure 9). Futterer

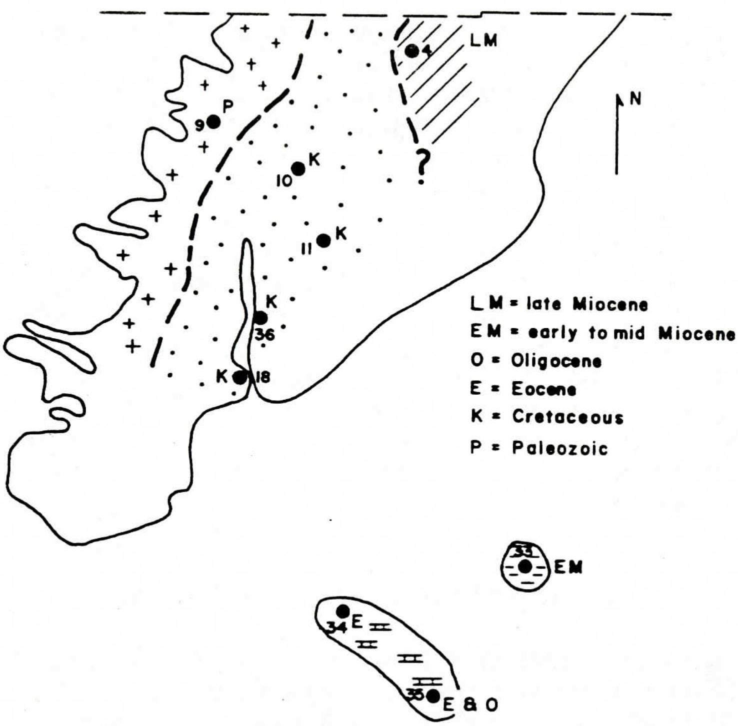


Figure 7. Subcrop map for Yorktown Formation in northeastern North Carolina. Main outcrop area bounded by solid line is same as given in Figure 1.

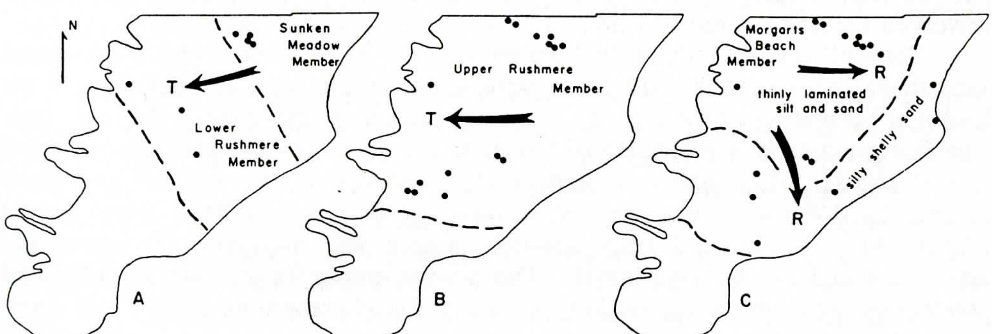


Figure 8. Yorktown facies maps for early Yorktown (*Pterygocythereis inexpectata* zone), A; middle Yorktown (*Orioina vaughani* zone), B; and late Yorktown (upper *O. vaughani* zone), C. Heavy arrows indicate general directions of trangression (T) and regression or progradation (R). Yorktown strata are undifferentiated in southern part of outcrop area. Dots indicate section localities used in Figure 1.

(1982) and other workers have noted that this fabric is often due to shell sorting by waves in very shallow water. Newell and Rader (1982) suggested that the Eastover - Yorktown contact is conformable downdip (in the lower James River area) but unconformable in updip areas. I think that this relationship is unlikely given the clear evidence for a hiatus in downdip areas in Virginia and North Carolina. It is probable that the pre-Yorktown hiatus is not everywhere of equal duration. The downdip or offshore hiatus, if produced primarily by submarine processes, might be of a smaller magnitude than the same break in the western part of the outcrop area.

Later in Yorktown history, and possibly after a pause in the intial transgression, or a decrease in the initially higher rate of sea level rise, a rapid southward and westward movement of the shoreline caused basal strata of the Rushmere Member to onlap progressively older pre-Yorktown beds. During the deposition of the upper Rushmere Member (lower *Orionina vaughani* assemblage zone), a shallow shelf sea extended to and slightly beyond the present Fall Line (Figure 8b and 10). During the upper *O. vaughani* assemblage zone an extensive mud facies prograded or regressed over most of the Yorktown basin (Figure 8C and 10). To the southwest these very silty, thinly laminated sediments of the Morgarts Beach Member grade into sandier sediments, and downdip they grade into an open shelf, presumably deeper water, silty sand facies with very high molluscan diversity.

PALEOENVIRONMENTAL SYNTHESIS

The depositional environment represented by most outcropping Yorktown facies in North Carolina is a shallow to moderately deep shelf with estimated depths of 10 to 40 m. Lagoonal or possible prodeltaic deposits are also present, but nearshore and very shallow water facies are uncommon. Sediment textures and sedimentary structures indicate that the basin floor was generally below normal (fair weather) wave base. Sediments and fossils were periodically reworked by storm surge and tidal currents.

The well sorted sands of the Sunken Meadow Member comprise a winnowed sand sheet that was produced by reworking underlying relict shelf sediments of the Eastover Formation and in part by shoreface erosion of older strata (Figure 10A). The fine grain sand is indicative of moderate energy conditions and homogeneous source strata. Thin layers of horizontal, often convex-up, pectens and other disarticulated bivalves may represent shell lag deposits of brief storm events (Figure 9). Thick shell beds are not present and mollusks are commonly articulated and evenly distributed. The diverse mollusks and the foraminiferal assemblage (Gibson, 1967; 1983) suggest a mid-shelf environment at a maximum depth of about 40 m.

The Rushmere Member represents mid to shallow shelf environments (10 to 40 m deep; Gibson, 1967; 1983). Thick shell beds are probably indicative of high molluscan productivity, more frequent bottom disturbance by storms, and other ecological factors. The presence of coarse sand, pebbles, and bone in some of these shell beds indicates that periods of slow deposition and storm winnowing were important in producing broad thin shell lenses (Figure 9). The unabraded, unbroken, and often articulated bivalves are evidence of mild current disturbance and limited transport of fossils. Storm wave pressure (Gernant, 1970) and taphonomic feedback (Kidwell, 1986) have been used to explain the origin of thick shell beds in the Miocene Choptank Formation in the Chesapeake Group of Maryland. Most shell beds in the Yorktown seem to require explanations that combine physical with ecological and biological processes.

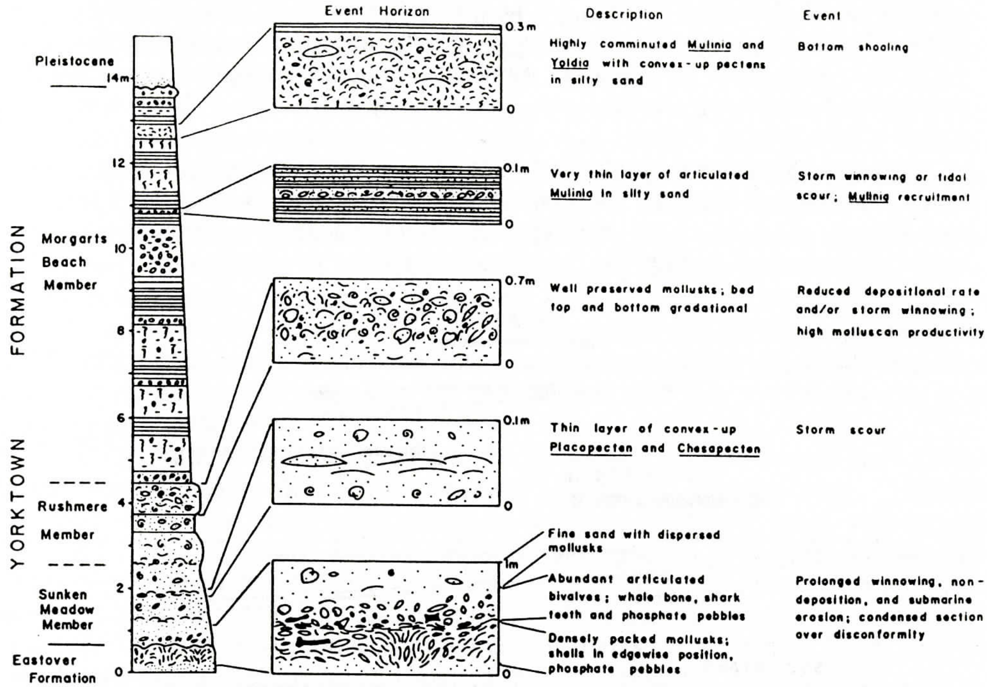


Figure 9. Types of stratigraphic event horizons in the Yorktown Formation. The composite section is for strata along the Meherrin River (localities 3 - 7) in the vicinity of Murfreesboro, N. C.

The thick sequence of thinly laminated muddy sediments of the Morgarts Beach Member requires low energy conditions and a substantial mud source. Massive coquina shoals or offshore bars (Johnson, 1969) formed in southern Virginia at about this time or very slightly later. A nearshore or prodeltaic mud blanket can be deposited in a bar-protected lagoonal environment or on a semi-protected shelf. Curray (1969) suggested that a mud blanket may be typical of a broad equilibrium shelf and should occur above basal transgressive sands. The fact that such a mud blanket is not currently found on the Atlantic shelf of the eastern United States may be attributed in part to the sediment trapping abilities of recently inundated and as yet unfilled estuaries (Curray, 1969). Once these coastal sediment traps are filled, mud will be bypassed to the shelf in quantity where some of it may accumulate as fine sediment blankets (Swift, 1969; McCave, 1972; Meade, 1972). This concept has significance for Yorktown deposition. The Sunken Meadow and Rushmere Members represent basal transgressive sands. Once Yorktown estuarine basins were filled and deltas formed, large amounts of fine sand, silt, and clay was bypassed (Figure 10) to the shelf (Newell and Rader, 1982). This sudden increase in direct supply accounts for the fact that at some outcrops the Morgarts Beach Member is more than twice as thick as the Sunken Meadow and Rushmere Members combined (Figure 4). The thinly laminated silt facies of the Morgarts Beach Member is thickest in the area of the Meherrin River

Valley. The fine sand lenses and laminations within the silt were produced by active bottom currents due to wave and tidal action. Episodic current action winnowed silt and clay to produce sand-rich lamina. At locality 13 small scale ripples document bottom sand transport. Storm currents created *Mulinia* shell-lag deposits 1-2 cm thick and local shoaling was associated with the accumulation of comminuted shells of *Mulinia* and *Yoldia* (Figure 9). That these occasionally reworked muds were at times accumulating under anoxic or dysaerobic conditions is indicated by common small pyrite crystals. Deposition of abundant organic matter is also suggested by the great numbers of deposit feeding bivalves *Yoldia* and *Nucula*. Details of the *Mulinia* and *Yoldia* distribution in the Yorktown and origin of their shell beds is the subject of a paper in preparation.

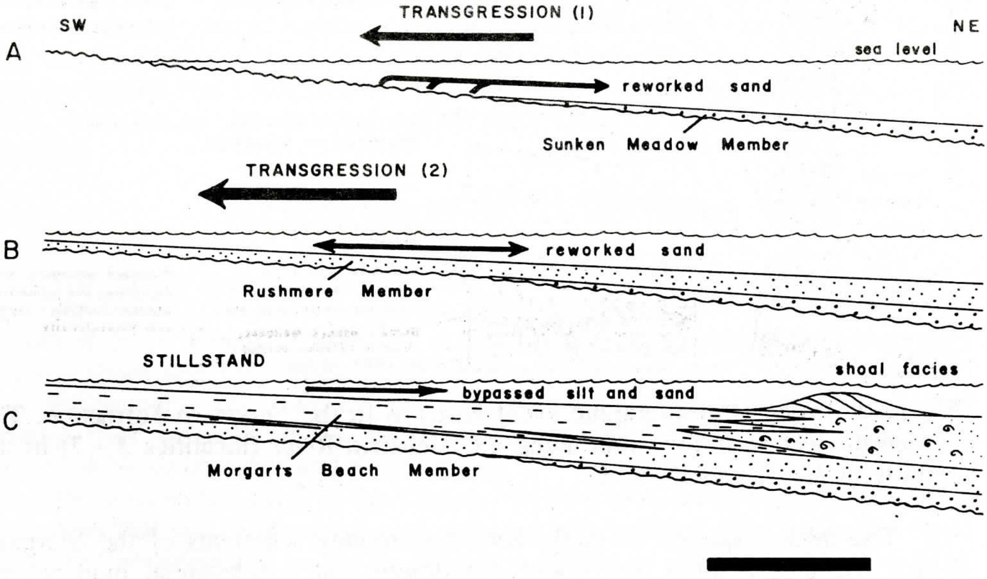


Figure 10. Yorktown depositional history in northeastern North Carolina. Lower *Pterygocythereis inexpectata* zone, A; upper *P. inexpectata* - mid *O. vaughani* zone, B; mid - upper *O. vaughani* zone, C. Shoal facies (not shown to scale) is exposed in outcrop in southeastern Virginia but not in northeastern North Carolina.

SUMMARY

The outcropping Yorktown Formation was deposited on a shallow to intermediate shelf during a late Pliocene asymmetric transgressive and regressive cycle. During the rapid and possibly pulsating transgression, a fine sand sheet comprised of the Sunken Meadow and Rushmere Members was reworked and winnowed by periodic storms and shelf currents. After the maximum extent of the transgression was reached, muds and very fine sands bypassed coastal environments to accumulate as an open lagoonal or prodeltaic shelf mud blanket. This blanket was commonly reworked by wave and tidal currents and occasionally by storm surge currents to produce interbedded sand laminae and shell beds

(Figure 9). The muddy sediments of the Morgarts Beach Member record stabilization of sea level and gradual regression. A full regressive facies sequence is not preserved because sea level dropped rapidly before significant shoreface, bar, and estuarine sediments could accumulate and because these facies in North Carolina were largely removed by latest Pliocene and early Pleistocene erosion.

Yorktown facies patterns and ultimate shoreline positions were controlled primarily by glacioeustatic changes in sea level (Blackwelder, 1981b). Differential subsidence of portions of the shelf created local basins, such as the Albemarle Embayment, which controlled regional thickness distribution and, to some degree, facies patterns. Given the relatively short time span represented by Yorktown strata, about 1 to 2 my (Blackwelder, 1981b), and the slow rate of subsidence in updip areas of the continental shelf (Pittman, 1978), variations in sea level and sediment supply were the primary controls of facies distributions. Sediment supply varied during the transgression/regression cycle, with the greatest rate of sedimentation occurring in the period after the maximum extent of transgression.

Differential subsidence produced the irregularly warped coastal plain surface that was flooded during the earliest Yorktown transgression. Modest sedimentation was intially localized by this pre-Yorktown basin architecture; however, later Yorktown facies completely onlapped all irregularities during the time of maximum sea level rise. After the transgressive maximum, increased rates of sedimentation caused progradation of muddy facies. Rapid sea level fall at the end of Yorktown time produced a thin cap of regressive nearshore facies that has been largely removed in North Carolina by post-Yorktown erosion.

The asymmetry of transgressive and regressive cycles, typical of many Neogene rock units of the mid-Atlantic Coastal Plain, is characteristic of a depositional system on a stable platform margin that is dominated by variations in sea level. In this case, sea level position rather than tectonism also indirectly controlled sediment supply. Sea level changes during Yorktown deposition are approximately equivalent to those noted for sedimentary sequences TB3.5 and TB3.6 as defined by Haq and others (1987). Differences in the Yorktown sea level curve (Figure 5) and in the relative amounts of coastal onlap when compared to the worldwide curve may reflect fourth order sea level perturbations not recorded by Haq and others (1987). These minor sea level changes noted for the Yorktown may be a result of local basin tectonism but it is possible that they are real eustatic sea level fluctuations that are simply beyond the limit of resolution of the presently published curve (Haq and others, 1987).

ACKNOWLEDGEMENTS

I would like to thank Walter H. Wheeler for introducing me to the Coastal Plain stratigraphy of North Carolina and for being a most patient advisor in the early history of this project. Warren C. Blow of the U.S. National Museum accompanied or led me into the field in Virginia and North Carolina on numerous occasions over the past 20 years. Druid Wilson provided advice on stratigraphy and molluscan identifications. Joseph E. Hazel, while with the U.S. Geological Survey, freely discussed and provided information on Yorktown ostracode biostratigraphy and he kindly identified ostracodes from key sections and placed

them in his zonal scheme. Lauck W. Ward and William Miller III reviewed the manuscript and provided many very helpful suggestions for its improvement.

LITERATURE CITED

Bailey, R. H., 1973, Paleoenvironment, paleoecology, and stratigraphy of molluscan assemblages from the Yorktown Formation (upper Miocene - lower Pliocene) of North Carolina: unpublished Ph.D. dissertation, University of North Carolina, Chapel Hill, 110p.

Bick, K. F. and Coch, N. K., 1969, Geology of the Williamsburg, Hog Island, and Bacons Castle quadrangles, Virginia: Virginia Division of Mineral Resources Report of Investigations 18, 28 p.

Blackwelder, B. W., 1981a, Stratigraphy of upper Pliocene and lower Pleistocene marine and estuarine deposits of northeastern North Carolina and southeastern Virginia: U.S. Geolological Survey Bulletin 1502 - B, 16 p.

_____, 1981b, Late Cenozoic stages and molluscan zones of the U.S. middle Atlantic Coastal Plain, Journal of Paleontology, v. 55, Memoir 12, 34 p.

_____, and Ward, L. W., 1979, Stratigraphic revision of the Pliocene deposits of North and South Carolina: South Carolina Geological Survey Geologic Notes, V. 79, p. 33-49.

Brown, P. M., 1985, Geologic map of North Carolina: North Carolina Department of Natural Resources and Community Development, Raleigh, N.C.

Curray, J. R., 1969, History of continental shelves, lecture 6, p. JC-VI1-18, *in* Stanley, D. J., ed., The New Concepts of Continental Margin Sedimentation, AGI short course lecture notes, Washington, D.C.

Futterer, E., 1982, Experiments on the distinction of wave and current influenced shell accumulations, p. 176-179, *in* Eisele, G. and Seilacher, A., eds., Cyclic and Event Stratification, Springer - Verlag, New York, 536 p.

Gernant, R. E., 1970, Paleoecology of the Choptank Formation (Miocene) of Maryland and Virginia, Maryland Geological Survey, Report of Investigations, No. 12, 90 p.

Gibson, T. G., 1967, Stratigraphy and paleontology of the phosphatic Miocene strata of North Carolina: Geololgical Society of America Bulletin, v. 78, p. 631-650.

_____, 1970, Late Mesozoic-Cenozoic tectonic aspects of the Atlantic coastal margin: Geological Society of America Bulletin, v. 81, p. 1813-1822.

_____, 1983, Stratigraphy of Miocene through lower Pleistocene strata of the United States Central Atlantic Coastal Plain, p. 35-80, *in* Ray, C. E., Geology and Paleontolgy of the Lee Creek Mine, North Carolina, I, Smithsonian Contributions to Paleobiology, No. 53.

Haq, B. U., Hardenbol, J., and Vail, P. R., 1987, Chronology of fluctuating sea levels since the Triassic: Science, v. 235, p. 1156-1167.

Hazel, J. E., 1971, Ostracode biostratigraphy of the Yorktown Formation (upper Miocene and lower Pliocene) of Virginia and North Carolina: U.S. Geological Survey Professional Paper 704, 13, p.

_____, 1977, Distribution of some biostratigraphically diagnostic ostracodes in the Pliocene and lower Pleistocene of Virginia and Northern Carolina: U.S. Geological Survey Journal of Research v. 5, p. 373-388.

_____, 1983, Age and correlation of the Yorktown (Pliocene) and Croatan (Pliocene and Pleistocene) Formations at the Lee Creek Mine, P. 81 - 199, *in* Ray, C. E., Geology and Paleontology of the Lee Creek Mine, North Carolina, I, Smithsonian Contributions to Paleobiology, No. 53.

Johnson, G. H., 1969, Guidebook to the geology of the York-James peninsula and south bank of the James River: College of William and Mary, Department of Geology Guidebook No. 1, 33p.

Kidwell, S. M., 1986, Taphonomic feedback in Miocene assemblages: testing the role of dead hardparts in benthic communities, Palaios, v. 1, p. 239-255.

Mansfield, W. C., 1929, The Chesapeake Miocene basin of sedimentation as expressed in the new geologic map of Virginia: Journal of Washington Academy of Sciences, v. 19, p. 263-268.

_____, 1944, Stratigraphy of the Miocene of Virginia and the Miocene and Pliocene of North Carolina, p. 1-19, *in* Gardner, J. A., Mollusca from the Miocene and lower Pliocene of Virginia and North Carolina: part 1, Pelecypoda: U.S. Geological Survey Professional Paper 199A.

McCave, I. N., 1972, Transport and escape of fine-grained sediment from shelf areas, p. 225-248, *in* Swift, D.J.P., Duane, D.B., and Pilkey, O.H., *eds*., Shelf Sediment Transport, Dowden, Hutchinson and Ross, Inc., Stroudsburg, PA.

Meade, R. H., 1972, Source and sinks of suspended matter on continental shelves, p. 249-279, *in* Swift, D.J.P., Duane, D.B., and Pilkey, O.H., *eds*. Shelf Sedient Transport, Dowden, Hutchinson and Ross, Inc., Stroudsburg, PA.

Newell, W. L., and Rader, E. K., 1982, Tectonic control of cyclic sedimentation in the Chesapeake Group of Virginia and Maryland, p. 1-27, *in* Lyttle, P.T., ed., Central Appalachian Geology, Geological Society of America field trip guidebook for NE-SE Mtg.

Pittman, W. C., 1978, Relationship between eustacy and stratigraphic sequences of passive margins, Geological Society of America Bulletin, v.89, p. 1389-1403.

Seilacher, A. 1982, General remarks about event deposits, p. 161-174, *in* Einsele, G. and Seilacher, A., eds., Cyclic and Event Stratification, Springer-Verlag, New York, 536 p.

Stanley, S. M., 1970, Relation of shell form to life habits in the Bivalvia (Mollusca): Geological Society America Memoir 125, 296 p.

Swift, D. J. P., 1969, Equilibrium versus disequilibrium shelves: evolution of a graded shelf, lecture 7, p. DS-7-1-19, *in* Stanley, D. J., ed., The New Concepts of Continental Margin Sedimentation, AGI short course lecture notes, Washington, D.C.

Ward, L. W., and Blackwelder, B. W., 1980, Stratigraphic revision of upper Miocene and lower Pliocene beds of the Chesapeake Group, middle Atlantic Coastal Plain, U.S. Geological Survey Bulletin 1982 - D, 61 p.

_____, and Strickland, G. L., 1985, Outline of Tertiary stratigraphy and depositional history of the U.S. Atlantic Coastal Plain, p. 87-123, *in* Poag, C. W., ed., Geologic Evolution of the United States Continental Margin, Van Nostrand Reinhold Company, Inc., New York.

APPENDIX

Location information for localities given in Figure 1.

1. North bank of Meherrin River, Northampton Co., 7.0 km NW of Severn, at bridge where N.C. County Road 1339 crosses river.
2. South bank of Meherrin River, Northampton Co., 6.8 km NW of Severn.
3. South Bank of Kirbys Creek, Northampton Co., 6.5 km SE of Severn, 30 m downstream from bridge where N.C. County Road 1362 crosses creek.
4. South bank of Meherrin River, Hertford Co. 2.2 km N of Murfreesboro, 3.0 upstream from U.S. Route 258 bridge across Meherrin.
5. South bank of Meherrin River, Hertford Co., Murfreesboro, 0.8 km upstream from U.S. Route 258 bridge across Meherrin.
6. South bank of Meherrin River, Hertford Co., Murfreesboro, 0.5 km upstream from U.S. Route 258 bridge across Meherrin.
7. South bank of Meherrin River, Hertford Co., 2.0 km E of Murfreesboro, 1.2 km downstream from U.S. Route 258 bridge across Meherrin.
8. Lower part of roadcut on N side of U.S. Route 158, Hertford Co., 20 W of bridge where U.S. 158 crosses Potecasi Creek.
9. Banks of northward flowing tributary to Quankey Creek, Halifax Co., N side of N.C. Route 561, 0.7 km W of railroad crossing over Route 561 in Halifax.
10. South bank of Roanoke River, Halifax Co., under bridge where U.S. Route 258 crosses Roanoke.
11. South bank of Roanoke River, Halifax Co., 50 m upstream from landing 1.3 km east of Palmyra.
12. South bank of Roanoke River, Halifax Co. 1.9 km downstream from landing 1.3 km east of Palmyra.
13. South bank of Roanoke River, Martin Co., 0.3 km downstream from public boat ramp 0.6 km east of Hamilton.
14. East bank of northward flowing tributary to Conoho Creek, Martin Co., 7.0 km NW of Williamston, outcrop is at end of concrete wall 50 m N of bridge where N.C. Route 125 crosses tributary.
15. Bank of small northward flowing creek, Martin Co., 3.2 km W of Williamston on N.C. Route 125, locality covered since 1970.
16. West bank of Fishing Creek, Edgecombe Co., 150 m downstream from bridge where N.C. Route 97 crosses creek.
17. South bank of Tar River, Edgecombe Co., 75 m upstream from bridge where N.C. Route 44 crosses Tar, 2.5 km N of Tarboro.
18. South bank of Tar River, Edgecombe Co., 0.5 km downstream from bridge where N.C. Route 42 crosses Tar, 0.6 km SE of Old Sparta.
19. South bank of Green Run, Pitt Co., under bridge where Rock Spring Road crosses Green Run, 150 m S of intersection of Rock Spring Road and Tenth Street, Greenville.
20. South bank of Tar River, Pitt Co., under bridge where N.C. County Road 1565 crosses Tar, 1.8 km NE of Grimesland.
21. South bank of Wiccacon Creek, Hertford Co., 0.1 km upstream from landing

where N.C. County Road 1433 terminates at creek.

22. West bank of Chowan River, Bertie Co., 1.8 km SE of Colerain, 200 m S from southernmost landing at Colerain beach.
23. West bank of Chowan River, Bertie Co., 2.0 km S of Colerain, 0.6 km S of southernmost landing at Colerain beach.
24. West bank of Chowan River, Bertie Co., 1.5 km NE of Mt. Gould, 0.4 km N of landing at end of N.C. County Road 1354.
25. West bank of Chowan River, Bertie Co., 2.0 km E of Mt. Gould, 0.4 km S of landing at end of N.C. County Road 1354.
26. West bank of Chowan River, Bertie Co., 0.3 km S of end of N.C. County Road 1360.
27. West bank of Chowan River, Bertie Co., 2.0 km N of bridge where U.S. Route 17 crosses Chowan.
28. West bank of Chowan River, Bertie Co., 1.3 km N of bridge where U.S. Route 17 crosses Chowan.
29. South bank of Tar River, Edgecombe Co., about 1.5 km upriver from northeasterly flowing creek (Buck Swamp) entering Tar River, about 14 km NW of bridge where N.C. Route 44 crosses Tar.
30. South bank of Tar River, Edgecombe Co., 12.5 km W of bridge where N.C. Route 44 crosses Tar.
31. South bank of Tar River, Edgecombe Co., 10.5 km NW of bridge where N.C. Route 44 crosses Tar.
32. South bank of Tar River, Pitt Co., about 8.8 km E (downriver) from Greenville.
33. Texasgulf Inc. open pit mine, Beaufort Co., about 12 km NE of Aurora, S bank of Pamlico sound.
34. Abandoned borrow pit, Craven Co., 150 m S of Neuse River, 5 m E of county road, about 3.0 km NE of Fort Barnwell.
35. Martin Marietta Co. Quarry, Craven Co., Newbern, 0.8 km NE of N.C. Route 55 on N.C. County Road 1402, upper bench at western end of quarry.
36. Borehole NW of Martin Marietta Co. Quarry, Pitt Co., Fountain, about 27 km NW of Greenville (Blackwelder, 1981).
37. South bank of Tar River, Edgecombe Co., about 4.0 km NE of Tarboro, at sharp bend in river about 4.5 km downriver from Seaboard Coast Line railroad bridge.

CONODONT BIOSTRATIGRAPHY OF THE BRUSH CREEK SHALE AND AMES SHALE UNITS OF THE GLENSHAW FORMATION IN THE MARYLAND COAL FIELDS AND OF CORRELATIVE STRATA, APPALACHIAN BASIN

GLEN K. MERRILL *University of Houston-Downtown, Houston TX 77002*

PAUL C. LYONS *U.S. Geological Survey, 956 National Center, Reston VA 22092*

ABSTRACT

A study of conodonts in the Upper Pennsylvanian Glenshaw Formation (Conemaugh Group) of the western Maryland coal fields allows correlation of the Ames shale unit of Maryland with the Ames Limestone Member and equivalent strata of Pennsylvania, West Virginia, Kentucky, and Ohio. This correlation is indicated by the almost total dominance in the Ames of *Streptognathodus elegantulus*. This species is typical of an offshore conodont biofacies. The conodont fauna of the underlying Brush Creek shale unit of Maryland comprises *Streptognathodus oppletus*, a lesser percentage of *S. cancellosus*, *Cavusgnathus*, and possibly a few specimens of *Idiognathodus*. This assemblage is younger than that in the Lower Brush Creek limestone unit in southeastern Ohio (or the Brush Creek unit of northern Ohio) and is most similar to the conodont fauna of the "Upper" Brush Creek limestone unit in southern Ohio, northwestern West Virginia, and northeastern Kentucky. The Brush Creek shale unit of Maryland contains about 11 percent *Cavusgnathus*, compared with a lesser percentage in the Ames shale unit of Maryland. This content indicates that the Brush Creek represents a more brackish, more nearshore paleoenvironment than the Ames unit in this area.

The conodont data are consistent with an interpretation that the Ames Limestone and Ames shale unit represent a widespread single transgression of low diachroneity in the northern and central Appalachian basin. Also, these data provide biostratigraphic justification for considering the Conemaugh to be a group in Maryland and for subdividing it at the top of the Ames shale unit, or equivalent strata, into the Glenshaw and Casselman Formations, as has been done in the Conemaugh Group of Pennsylvania.

INTRODUCTION

The Conemaugh Group in the coal fields of Maryland (Fig. 1) contains in its lower half four marine or marginal marine units (Lyons and others, 1985) that have been referred to, from bottom to top, as the Brush Creek shale member of Hennen and Reger (1913), the Meyersdale limestone unit of Clark and others (1902), the Woods Run shale unit of Swartz and Baker (1922), and the Ames shale unit of Hennen and Reger (1913). Although Price (1913) considered the Ames and Brush Creek units in Maryland to be separable on the basis of the invertebrate mega-

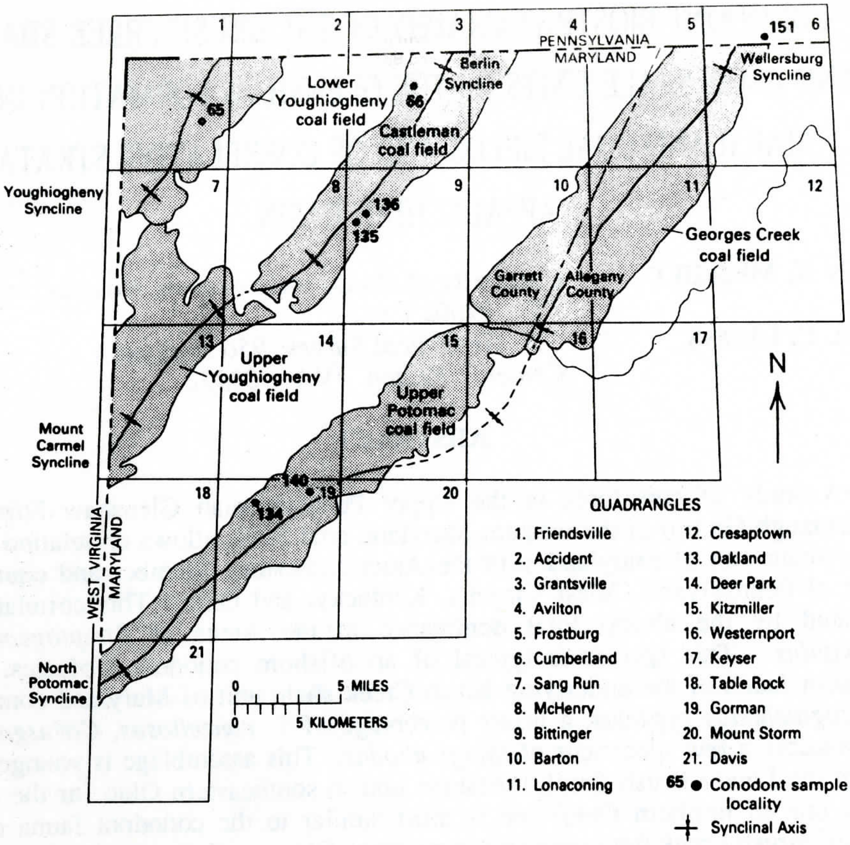


Figure 1. Coal fields of western Maryland showing the conodont sample localities.

fauna, this separation has not been verified by recent megafaunal collections (Lyons, unpub. data). We began our study of the conodont biostratigraphy in Maryland several years ago in an attempt to separate the Ames shale and Brush Creek shale units of Maryland and to determine their regional correlations. This paper reports on our investigations.

PHYSICAL STRATIGRAPHY

Figure 2 shows the generalized physical stratigraphy of the Pennsylvanian System in the Maryland coal fields. The Pennsylvanian sequence is about 2,000 ft (610 m) thick and contains five major lithostratigraphic units from bottom to top: Pottsville Formation, Allegheny Formation, Conemaugh Group (Glenshaw and Casselman Formations as recognized in Maryland in this paper), Monongahela Formation, and Waynesburg Formation (Lyons and Jacobsen, 1981). All of these strata are coal bearing and contain a total of more than 40 coal beds. The most economically important coal beds are the Lower, Middle, and Upper Kittanning, Upper Freeport, Lower Bakerstown, Harlem, Barton, Franklin, Pittsburgh, and Sewickley.

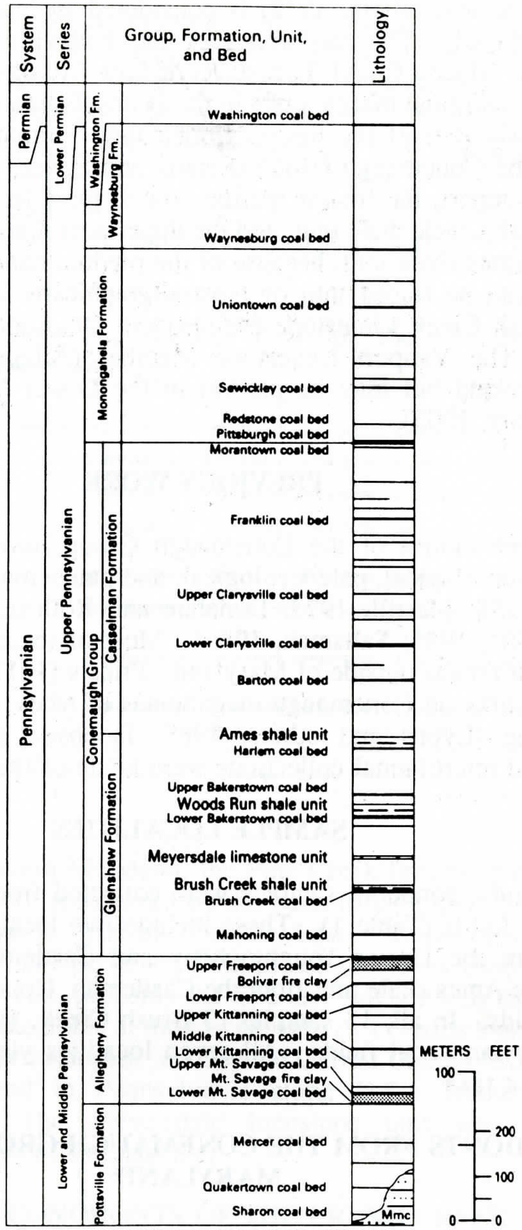


Figure 2. Generalized stratigraphy of the Pennsylvanian System in Maryland.

The most complete Middle and Upper Pennsylvanian sequence in the area is found in the Georges Creek coal field where erosion has removed the least amount of Upper Pennsylvanian strata. The Pottsville strata are thickest in the southern part of the Upper Potomac coal field, where their maximum thickness is about 500 ft. In most parts of Maryland, only the uppermost part of the Pottsville Formation and younger Pennsylvanian strata are present, and the entire Lower Pennsylvanian section (Englund and others, 1979) is considered in this report to be generally absent in Maryland. The sandstones of the Pottsville Formation rest unconformably on the Mauch Chunk Formation of Late Mississippian age.

Marine or marginal marine units in the Pennsylvanian of Maryland are known only in the lower part of the former Conemaugh Formation, now the Glenshaw Formation of the Conemaugh Group (Swartz and Baker, 1922; Lyons and others, 1985). In this report, the lowest marine unit of the Glenshaw Formation here is termed the Brush Creek shale unit, and the uppermost marine unit of this formation is termed the Ames shale unit, because of the predominance of shale lithology and because they can be traced into or biostratigraphically correlated with the type Ames and Brush Creek Limestone Members in Ohio and western Pennsylvania, respectively. The Vanport Limestone Member (Allegheny Formation) is not known in Maryland but may be present in the Lower Youghiogheny coal field (Clark and others, 1905).

PREVIOUS WORK

The marine faunas of the Conemaugh Group have been the subject of a number of paleontological, paleoecological, and paleoenvironmental papers (Price, 1913; Lintz, 1958; Merrill, 1973; Donahue and Rollins, 1974; Brezinski, 1983; Brady and others, 1985; Saltsman, 1986). Most of these papers have dealt with the Conemaugh faunas outside of Maryland. Price's (1913) and Lintz's (1958) are the principal works on Conemaugh megafaunas of Maryland. In connection with recent mapping (Lyons and others, 1985; Jacobsen and Lyons, 1985), new megafaunal and microfaunal collections were made of the marine Glenshaw units.

SAMPLE LOCALITIES

In this study, conodont samples were collected from seven localities in the Maryland coal fields (Table 1). These include two localities of the Brush Creek shale unit from the Lower Youghiogheny and Castleman coal fields and five localities of the Ames shale unit from the Castleman, Upper Potomac, and Georges Creek coal fields. In all, 13 samples (3 Brush Creek, 10 Ames) were processed from four Maryland coal fields. All seven localities yielded biostratigraphically useful conodont taxa.

CONODONTS FROM THE CONEMAUGH GROUP IN WESTERN MARYLAND

Conemaugh strata, mostly Glenshaw Formation (lower part of Conemaugh Group), in the central and northern Appalachian basin include thin, marine fossil-bearing beds within dominantly nonmarine successions. Generally, at least one marine unit is present in every area where Conemaugh strata occur, and commonly several units are present. Six or seven marine Conemaugh units are included along the most marine-influenced part of the outcrop in east-central Ohio; the number of units generally decreases eastward and southward toward the source of terrigenous sediments. Therefore, in the proposed Pennsylvanian Stratotype (Englund and others, 1979), marine units are not known in the Conemaugh. In most areas of Kentucky, West Virginia, and Maryland two marine units are present; in some places, three, and rarely four, marine units are present in the Conemaugh Group. Where two marine units are present, the upper usually is identified as the Ames Limestone Member, and the lower as the Brush Creek Limestone Member or equivalent strata. If two other marine units are present in between, they generally are identified as the Cambridge (Pine Creek) Limestone Member (Sturgeon and others, 1958) below and Portersville shale or limestone member (Condit, 1912; equivalent to Woods Run shale unit or Friendsville shale unit) above.

Table 1. Conodont sample and locality data for the Glenshaw Formation (Conemaugh Group) of Maryland.

Stratigraphic Unit	Coal Field*	Locality	Samples	UTM Location	Geographic Coordinates	Quadrangle	Description
Ames shale unit	UP	134	A,B,C	17SPP44255692	39°22'06"N 79°19'30"W	Gorman	1 mi. S of Kelso Gap
Ames shale unit	C	135	A,B	17SPP51588160	39°34'12"N 79°14'06"W	Bittinger	2 mi. S of Bittinger
Ames shale unit	C	136	-	17SPP51848207	39°34'39"N 79°13'56"W	Bittinger	near above on Maryland Route 495
Ames shale unit	UP	140	A,B	17SPP48225883	39°22'07"N 79°16'46"W	Gorman	near Bethlehem School
Ames shale unit	GC	151	A,B	17SPP84539939	39°43'35"N 78°50'49"W	Cumberland	borrow pit east of PA160, Wellersburg, PA
Brush Creek shale unit	LY	65	-	17SPP36149109	39°39'48"N 79°24'47"W	Friendsville	Friendsville exit on Interstate 48
Brush Creek shale unit	C	66	A,B	17SPP56289524	39°41'39"N 79°10'42"W	Grantsville	Interstate 48, about 1 mi. west of Route 495

* UP, Upper Potomac; C, Castleman; LY, Lower Youghiogheny; GC, Georges Creek.

-, one sample

In our study of western Maryland, the Pine Creek limestone or Friendsville shale marine units were not recognized on the basis of conodont data, despite the presence of the Friendsville type area in the Lower Youghiogheny coal field (Jacobsen and Lyons, 1985). However, field investigations in 1986 have revealed the presence of a marine unit, probably the Friendsville, between the Brush Creek shale and Woods Run shale units, at the Friendsville exit roadcuts on Interstate 48. Pine Creek strata have been sampled just beyond Garrett County in West Virginia, near Bruceton Mills, and in Pennsylvania along U.S. Route 40 near the Youghiogheny River. The Meyersdale limestone unit was also recently recognized by us just west of Friendsville.

STRATIGRAPHY AND CONODONTS OF THE BRUSH CREEK LIMESTONE AND SHALE UNITS OF THE CONEMAUGH AND GLENSHAW FORMATIONS

Strata of the so-called Brush Creek Limestone or shale unit of the Conemaugh or Glenshaw Formations are found in all five states where strata of the Conemaugh Group crop out. Distribution of these marine rocks is less widespread than that of the Ames Limestone Member and equivalent strata of the Conemaugh or Glenshaw Formations. In some areas, Brush Creek strata have not been found. In contrast, in parts of southeastern Ohio from Gallia County (Condit, 1912) to

Athens County (Sturgeon and others, 1958) and in northeastern Kentucky (Spencer, 1964), two units nearly 10 m apart are called "Lower" and "Upper" Brush Creek limestone of the Conemaugh Formation. In these areas, the conodont faunas of the two units, although similar, are distinguishable (Merrill, 1964, 1974; Lane and others, 1971). Where a single Brush Creek horizon occurs, as in western Maryland, the range of the ages of conodont faunas is at least equivalent to the range of the ages of conodonts in the Brush Creek strata of southeastern Ohio (Merrill 1964, 1974; Lane and others, 1971) and of northeastern Kentucky (Merrill, unpub. data) and may be greater. The range in ages among so- called Brush Creek strata is considerable.

The samples from western Maryland (Table 1), like other Brush Creek samples, are dominated by forms transitional between *Idiognathodus* and *Streptognathodus* but relatively well advanced along this phylogenetic gradient. By using Ellison's (1941) form-taxonomy, these would be mostly *Streptognathodus oppletus* (Figure 3A) with some of the less advanced *S. cancellosus* (Figures 3B and 3C) and the more advanced *S. elegantulus* (Figures 3D and 3E). A few forms can assigned to *Idiognathodus*. This microfaunal assemblage is most similar to that of the "Upper" Brush Creek Limestone Member in southern Ohio and the Brush Creek Limestone Member in parts of northwestern West Virginia and northeastern Kentucky (Merrill, 1964, 1974; Lane and others, 1971). The assemblage clearly is younger than the faunas dominated by *Streptognathodus cancellosus* and *Idiognathodus delicatus* in the "Lower" Brush Creek Limestone (of Condit, 1912; and of Sturgeon and others, 1958) strata in southeastern Ohio or the Brush Creek Limestone Member of Stout and Lamborn, 1924, in northern Ohio.

Conodont biotas in the Brush Creek strata in western Maryland average about 11 percent *Cavusgnathus*. These biotas reflect some influx of fresh water but not extremely brackish conditions. These environments were more influenced by freshwater than those of the Ames Limestone and shale units of the Glenshaw Formation of Maryland, and so they probably represent more nearshore deposits.

AREAL EXTENT AND CONODONT FAUNA OF THE AMES LIMESTONE MEMBER OF THE GLENSHAW FORMATION

The Ames Limestone Member of the Glenshaw Formation (Conemaugh Group) is the most widespread marine unit in the Pennsylvanian of the Appalachians (Flint, 1965). The area underlain by the Ames Limestone Member and equivalent strata is at least 53,000 km^2 and originally was probably considerably more. Most older marine units of the Conemaugh Group in the region are very restricted laterally and generally have a single name applied to several stratigraphic units that represent different ages and genetically unrelated transgressive-regressive pulses. In contrast, the Ames appears to be an integrated stratigraphic unit deposited during a single transgressive and regressive episode. Although any such extensive stratigraphic unit must be diachronous, the diachroneity in the Ames is below the current level of conodont biostratigraphic resolution. This low diachroneity stems in part from the reduced taxonomic change undergone by Late Pennsylvanian (Virgilian) conodont faunas and probably, in part, from the Ames transgression having been very rapid and widespread. As a result, conodont faunas from Ames strata are generally very similar wherever they are collected within the Appalachian basin. They are also of low diversity.

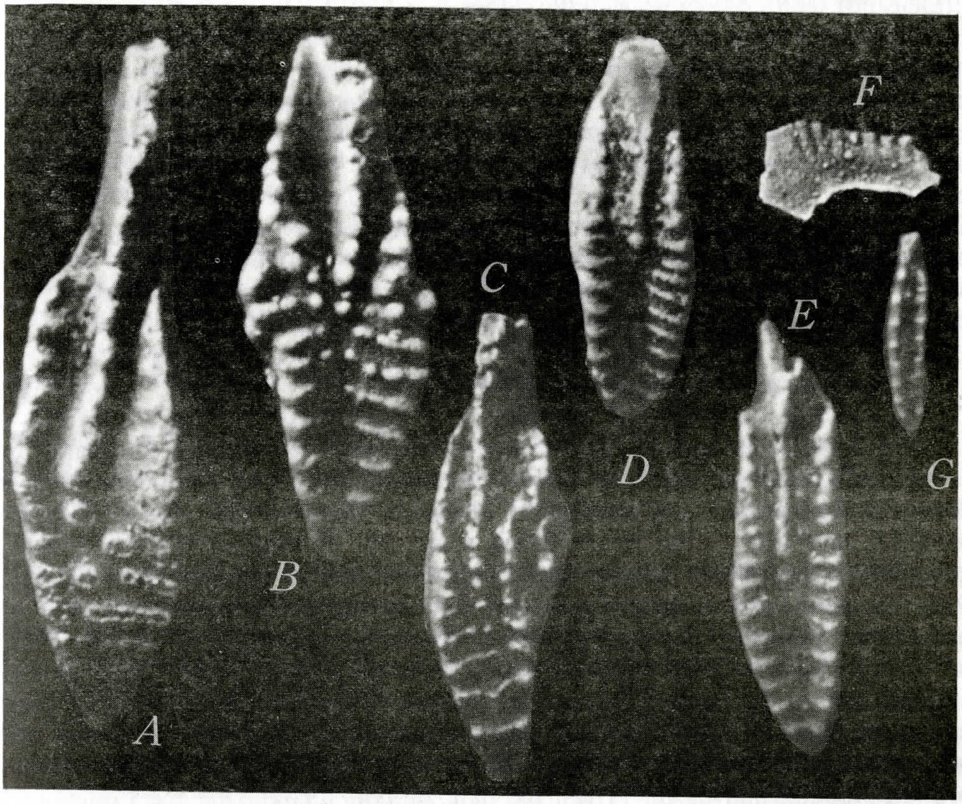


Figure 3. Selected conodont species occurring in the Brush Creek and Ames shale units of Maryland. A: *Streptognathodus oppletus* Ellison (Pa element, Brush Creek loc. 66B, Royal Ontario Museum Number (ROM) no. 45090); B, C: *Streptognathodus cancellosus* (Gunnell) (Pa elements, Brush Creek loc. 66B, ROM nos. 45091 and 45092); D, E: *Streptognathodus elegantulus* Stauffer and Plummer (Pa elements, Ames loc. 135B, ROM nos. 45093 and 45094); F: *Hindeodus ellisoni* (Merrill) (Pa element, Ames loc. 135B, ROM no. 45095); G: *Cavusgnathus merrilli* von Bitter (sinistral Pa element, Ames loc. 135B, ROM 45096). All illustrations 80X.

Two "species" of *Streptognathodus* dominate Ames strata, *S. elegantulus* (Figures 3D and E) and *S. elongatus* (possibly juveniles of *S. elegantulus*). Rarely, such species as *S. simulator* or *S. wabaunsensis* are found, but the dominance by *S. elegantulus* is nearly 100% for Ames strata (Lane and others, 1971). Also, representatives of this species from the Ames are biometrically distinguishable from older representatives of the species from stratigraphically lower units in the Conemaugh Group by the substantially shorter carina-to-platform relationship in adult specimens (Merrill and Wentland, 1986).

The low-diversity conodont fauna is a significant factor in identifying the

Ames Limestone Member and correlative strata. In addition to the streptognathodontids, few taxa other than representatives of *Hindeodus* (Figure 3F), *Idioprioniodus*, *Aethotaxis*, and *Diplognathodus* (which occurs only in northeastern Ohio) are present in the Ames.

Through a large part of the Ames area, the conodont biotas are dominated by an offshore *Streptognathodus* biofacies. As the ancient shoreline is approached near Huntington, West Virginia, and elsewhere toward the south and east, the nearshore, abnormal salinity-influenced *Cavusgnathus* biofacies increases in statistical importance. Within a few kilometers of the ancient shoreline, the percentage of *Cavusgnathus* species approaches 100 percent. None of the Ames samples studied from western Maryland reaches the 10 percent *Cavusgnathus* level considered to be the lower threshold of the *Cavusgnathus* biofacies. On this basis, the western Maryland Ames shale unit appears to be a more offshore deposit than the Brush Creek shale unit of the Glenshaw Formation that preceded it.

CONCLUSIONS

The Ames shale unit of the Glenshaw Formation in Maryland correlates, on the basis of conodonts, with the Ames Limestone Member of the Conemaugh or Glenshaw Formations and equivalent strata of Pennsylvania, West Virginia, Kentucky, and Ohio. The Brush Creek shale unit of Maryland is more nearly equivalent to the "Upper" Brush Creek Limestone unit in southern Ohio (Condit, 1912; Sturgeon and others, 1958) and the Brush Creek Limestone unit of northeastern Kentucky (Spencer, 1964; Connor and Flores, 1978; Merrill, 1986) and of northwestern West Virginia than it is to the Lower Brush Creek unit of southern Ohio or the Brush Creek unit of northern Ohio (Merrill, 1964, 1974; Lane and others, 1971). The conodont data indicate that the Ames Limestone and Ames shale unit represent a single widespread transgression in the central and northern Appalachian basin. Thus, the data support recognizing the Conemaugh as a group in Maryland and subdividing it into the Glenshaw and Casselman Formations; the top of the Ames shale unit defines the contact between the Glenshaw and the Casselman Formations (Flint, 1965).

ACKNOWLEDGMENTS

The authors thank Edwin F. Jacobsen, formerly of the U.S. Geological Survey, for collecting the Brush Creek samples from the Lower Youghiogheny coal field and Carl B. Rexroad, Indiana Geological Survey, for taking the photomicrographs.

REFERENCES CITED

Brady, K.B.C., Harper, J.A., and Inners, J.D., 1985, Ames or not Ames? An ambiguous Conemaugh marine zone at Ebensburg: Pennsylvania Geology, v. 16, no. 6, p. 7-12.

Brezinski, D.K., 1983, Development model for an Appalachian Pennsylvanian marine incursion: Northeastern Geology, v. 5, no. 2, p. 92-99.

Clarke, W.B., Abbe, C., Jr., Martin, G.C., Dorsey, C.W., Fassig, O.L., Pressey,

H.A., Paul, E.G., Bauer, L.A., Curran, H.M., and Sudworth, G.B., 1902, Garrett County (report): Maryland Geological Survey, 340 p.
Clark, W.B., Martin, G.C., Rutledge, J.J., Randolph, B.S., Stockton, N.A., Penniman, W.B.D., and Browne, A., 1905, Report on the coals of Maryland: Maryland Geological Survey, 651 p.
Condit, D.D., 1912, Conemaugh Formation in Ohio: Ohio Division of Geological Survey, 4th ser., Bulletin 17, 363 p.
Connor, C.W., and Flores, R.M., 1978, Geologic map of the Louisa quadrangle, Kentucky-West Virginia: U.S. Geological Survey Geologic Quadrangle Map GQ-1462, scale 1:24,000.
Donahue, J., and Rollins, H.B., 1974, Paleoecological anatomy of a Conemaugh (Pennsylvanian) marine event: Geological Society of America Special Paper 148, p. 153-170.
Ellison, S.P., Jr., 1941, Revision of the Pennsylvanian conodonts: Journal of Paleontology, v. 15, p. 107-143.
Englund, K.J., Arndt, H.H., and Henry, T.W., 1979, Proposed Pennsylvanian System Stratotype, Virginia and West Virginia: The American Geological Institute, AGI Selected Guidebook Series, no. 1, 136 p.
Flint, N.K., 1965, Geology and Mineral resources of southern Somerset County, Pennsylvania: Pennsylvania Geological Survey, 4th ser., County Report C-56A, 267 p.
Hennen, R.V., and Reger, D.B., 1913, Marion, Monongalia, and Taylor Counties: Morgantown, West Virginia Geological Survey Report, 844 p.
Jacobsen, E.F., and Lyons, P.C., 1985, Coal geology of the Lower Youghiogheny coal field, Garrett County, Maryland: U.S. Geological Survey Coal Investigation Map C-101, with text.
Lane, H.R., Merrill, G.K., Straka, J.J., II, and Webster, G.D., 1971, North American Pennsylvanian conodont biostratigraphy, *in* Sweet, W.C., and Bergström, S.M., eds., Symposium on conodont biostratigraphy: Geological Society of America Memoir 127, p. 407- 412.
Lintz, J., Jr., 1958, The fauna of the Ames and Brush Creek shales of the Conemaugh Formation of western Maryland: Journal of Paleontology, v. 32, no. 1, p. 97-112.
Lyons, P.C., and Jacobsen, E.F., 1981, Sources of data, procedures, and bibliography for coal resources investigation of western Maryland: U.S. Geological Survey Open-File Report 81-739, 33 p.
Lyons, P.C., Jacobsen, E.F., and Scott, B.K., 1985, Coal geology of the Castleman coal field: U.S. Geological Survey Coal Investigations Map C-98, with text.
Merrill, G.K., 1964, Zonation of platform conodont genera in Conemaugh strata of Ohio and vicinity: University of Texas, Austin, unpublished M.A. thesis, 175 p.
Merrill, G.K.,1973, Pennsylvanian conodont paleoecology: Geological Society of America Special Paper 141, p. 239-274.
Merrill, G.K., 1974, Pennsylvanian conodont localities in northeastern Ohio: Ohio Division Geological Survey, Guidebook 3, 28 p.
Merrill, G.K., 1986, Lithostratigraphy and lithogenesis of Conemaugh (Carboniferous) depositional systems near Huntington, West Virginia: Southeastern Geology, v. 26, no. 3, p. 155-171.

Merrill, G.K., and Wentland, E.M., 1986, Morphometrics in the group of *Streptognathodus elegantulus* (Conodonta, Carboniferous/Virgilian) for high-resolution biostratigraphy in the Appalachians and beyond: Geological Society of America, Abstracts with Programs, v. 18, p. 255.

Price, W.A., 1913, The invertebrate fauna of Maryland: Baltimore, Md; Johns Hopkins University, unpublished Ph.D. dissertation, 390 p.

Saltsman, A.L., 1986, Paleoenvrionments of the Upper Pennsylvanian Ames Limestone and associated rocks near Pittsburgh, Pennsylvania: Geological Society of America Bulletin, v. 97, p. 222-231.

Spencer, F.D., 1964, Geology of the Boltsfork quadrangle and part of the Burnaugh quadrangle, Kentucky: U.S. Geological Survey Geologic Quadrangle Map GQ-316, scale 1:24,000.

Stout, Wilbur, and Lamborn, R.E., 1924, Geology of Columbiana County: Ohio Geological Survey Bulletin 28, 408 p.

Sturgeon, M.T., and others, 1958, The geology and mineral resources of Athens County, Ohio: Ohio Division of Geological Survey, Bulletin 57, 600 p.

Swartz, C.K., and Baker, W.A., 1922, The coal formations and mines of Maryland, *in* Second Report of the coals of Maryland: Maryland Geological Survey, v. 11, pt. 1, 296 p.

REGIONAL GEOLOGIC FRAMEWORK SUMMARY OF THE NEOGENE-QUATERNARY LOUISIANA CONTINENTAL SHELF, NORTHERN GULF OF MEXICO[1]

GERALD L. SHIDELER *U.S. Geological Survey, Denver, CO 80225*

ABSTRACT

The highly petroliferous Neogene-Quaternary deposits of the Louisiana continental shelf in the northern Gulf Coast Basin were analyzed to determine their regional structural and stratigraphic relationships. The study was based on subsurface data from oil and gas wells penetrating to depths as great as 19,000 ft (5,791 m) below sea level.

A regional network of dip and strike cross sections illustrates a gulfward dipping and thickening wedge of terrigenous siliciclastic deposits modified by a complex fabric of structural elements. The structures were formed mainly from intrabasinal geostatic stresses associated with extensive depositional loading, which resulted in widespread diapirism and gravity failure. Major structural elements include numerous systems of coast-parallel, syndepositional faults that have down-to-basin displacements, along which much of the gulfward sectional thickening occurs. Numerous post-depositional folds and gravity fault systems are common. Piercement salt domes also are pervasive shelf structures, some of which indicate an age of emplacement as young as late Pleistocene.

Chronostratigraphic units ranging in age from early Miocene (Burdigalian) to late Pleistocene (Wisconsinan) show regional variations in stratigraphic thickness that reflect both gulfward and coast-parallel migrations of the basinal depocenter. Regional lithofacies relations delineated by induction-electrical logs illustrate the presence of three coast-parallel magnafacies, as defined on the basis of sand-shale proportions, which are related to deltaic depositional systems. Gulfward facies changes to progressively more argillaceous units indicate a downdip transition from continental to deep-water marine paleoenvironments. The magnafacies distribution patterns are temporally variable. They indicate evolutionary changes in the gross morphology of the Gulf Basin's fluvial-dominated delta systems, as well as the gulfward progradation of the northern Gulf continental margin during Neogene-Quaternary time.

INTRODUCTION

The Neogene-Quaternary geologic framework of the Louisiana continental shelf in the northern Gulf of Mexico is of major importance, both in terms of its natural-resource potential and as the basis for documenting the late geologic evolution of the northern Gulf Coast Basin. The Louisiana continental shelf has long been the most prolific offshore hydrocarbon province in the nation. In terms of the average production from combined Federal Outer Continental Shelf (OCS)

[1]Approved for publication by the Director, U.S. Geological Survey

areas between 1954 and 1983, the Louisiana OCS accounted for 96 percent of the total national OCS oil production and 95 percent of the total national OCS gas production (Lynch and Rudolph, 1984). This prolific hydrocarbon production is mainly from Neogene-Quaternary deposits; consequently, their economic importance is readily apparent. The Neogene-Quaternary section of the Louisiana shelf also provides valuable documentation regarding the evolution of the northern Gulf of Mexico continental margin.

As a result of the extensive long-term hydrocarbon exploration and production activities on the Louisiana shelf, the drilling of many thousands of offshore oil and gas wells has generated a voluminous amount of subsurface data for geologic investigations. This well database was utilized in a regional study of the important Neogene-Quaternary section of the Louisiana shelf (Shideler, 1986); the study involved preparing a network of regional geologic cross sections illustrating the general structural and stratigraphic characteristics of Neogene-Quaternary deposits. The purpose of the present report is to provide a summary discussion of the major regional geologic relationships illustrated by the aforementioned cross-sectional network, and to examine these relationships within the context of the northern Gulf of Mexico continental margin evolution.

The Louisiana continental shelf study area is located in the northern Gulf of Mexico (Fig. 1); it extends seaward from the Louisiana coastline to a depth of approximately 60 fathoms (110 m) at the shelf break, which separates the shelf from the steeper continental-slope province. The study area extends eastward from the Texas-Louisiana boundary to just east of the modern Mississippi River Delta, and it emcompasses an area of approximately 26,000 square miles (67,340 km^2). The Louisiana shelf shows a substantial variation in width, ranging from a maximum of about 125 miles (200 km) at the western extremity to less than 10 miles (16 km) off the modern Mississippi Delta in the east. Physiographically, the most conspicuous shelf feature is the northwest-trending Mississippi Canyon. The origin of the canyon has generally been attributed to entrenchment by a Pleistocene ancestral Mississippi River system during glacio-eustatic low stands of sea level, or alternately, to massive shelf-edge slope failure (Coleman and others, 1982).

The Louisiana Continental shelf is part of the northern Gulf of Mexico Basin, which appears to have begun forming in Triassic time by divergent plate-tectonic processes along the trailing passive margin of the North American Plate. Much of our knowledge regarding the basin origin has been reviewed in a symposium volume (Pilger, 1980). Laramide tectonism during latest Cretaceous-early Tertiary time provided voluminous quantities of terrigenous siliciclastic sediment from uplifted source areas for deposition within the northern Gulf Basin throughout the Tertiary period. Voluminous sediment influx continued during the Pleistocene, partly controlled by the advances and recessions of continental glaciers in North America. This continued sediment influx resulted in the accumulations of a thick wedge of Cenozoic deposits, locally exceeding 50,000 feet (15 km) in aggregate thickness (Curtis and Picou, 1978). The basinal depocenter, located along successive shelf edges, migrated throughout Cenozoic time in response to shifting sediment sources (e.g. Hardin, 1962; Woodbury and others, 1973; Martin, 1978); depocenter migration occurred both in gulfward and lateral directions (Fig. 2). Because rapid sedimentation exceeded the rate of regional subsidence, the northern Gulf shelf edge prograded as much as 402 km

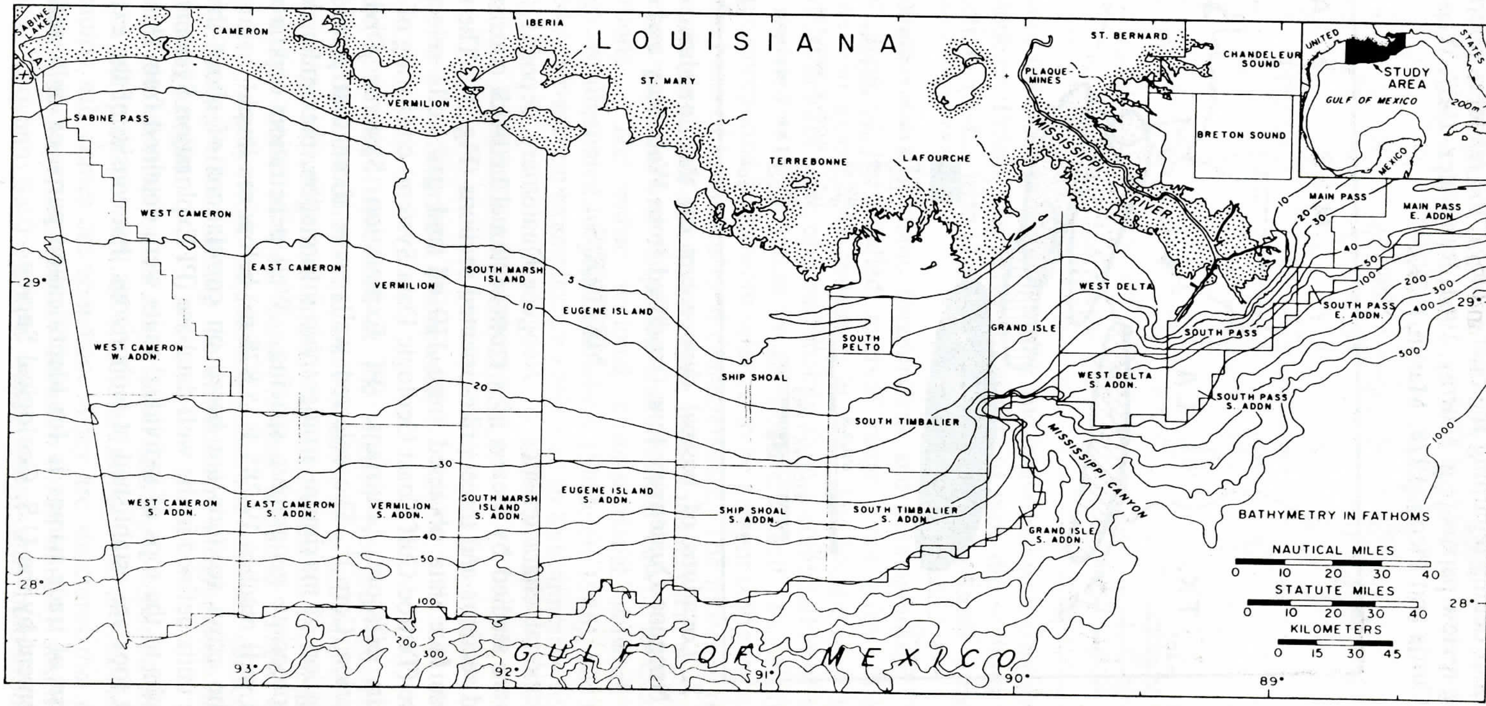


Figure 1. Location map and bathymetry of the Louisiana continental shelf study area. Geographic subdivisions of the shelf are U.S. Dept. of Interior central Gulf of Mexico OCS Areas.

(250 mi) gulfward since the end of the Cretaceous (Woodbury and others, 1973). For further details regarding regional geology, the reader is referred to the several existing review papers (e.g. Murray, 1961; Rainwater, 1967; Woodbury and others, 1973; Curtis and Picou, 1978; Martin 1978).

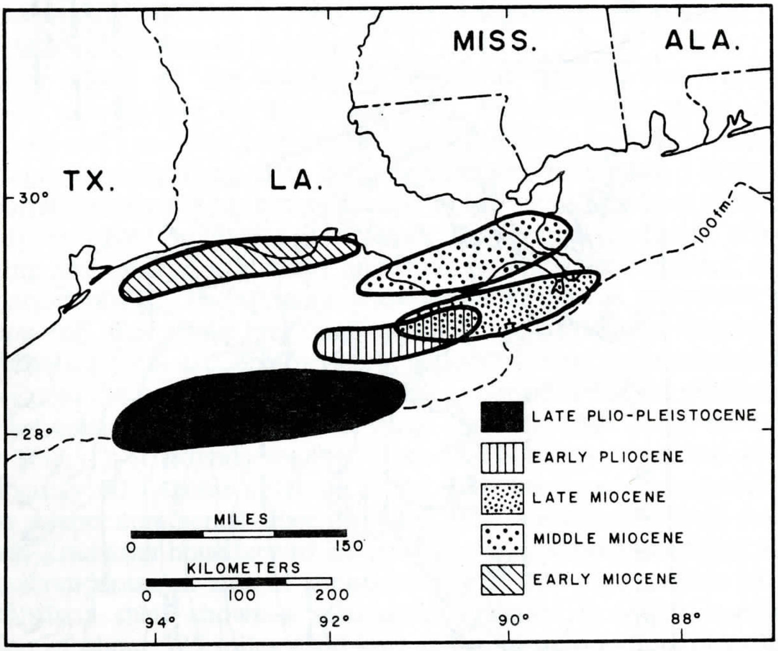


Figure 2. Locations of basinal depocenters on the Louisiana continental shelf during Neogene-Quaternary time (modified from Woodbury and others, 1973).

METHODS

Regional characteristics of Neogene-Quaternary deposits of the Louisiana shelf were studied by means of a cross-sectional network consisting of nine dip-oriented sections and three strike-oriented sections (Fig. 3). The network is based on subsurface data obtained from 110 oil and gas wells selected from a well database (Tenroc Gulf Coast Geologic Data System) consisting of several thousand wells in offshore Louisiana (PI Exploration Systems Division, Petroleum Information Corp.)[1]. The selected wells were those that provided an optimum combination of maximum stratigraphic information, maximum depth penetration, and appropriate geographic spacing. Well penetration depths are variable; the deepest well reaches 19,285 ft (5,878 m) below sea level.

The cross sections are based on correlations of about thirty chronostratigraphic units defined in the well database (PI Exploration Systems Division, 1985); the depths to the tops of individual units were obtained from this database, where the unit tops are established at biohorizons that consist of the extinction horizons

[1]The use of trade names is for identification purposes only and does not imply endorsement by the U.S. Geological Survey.

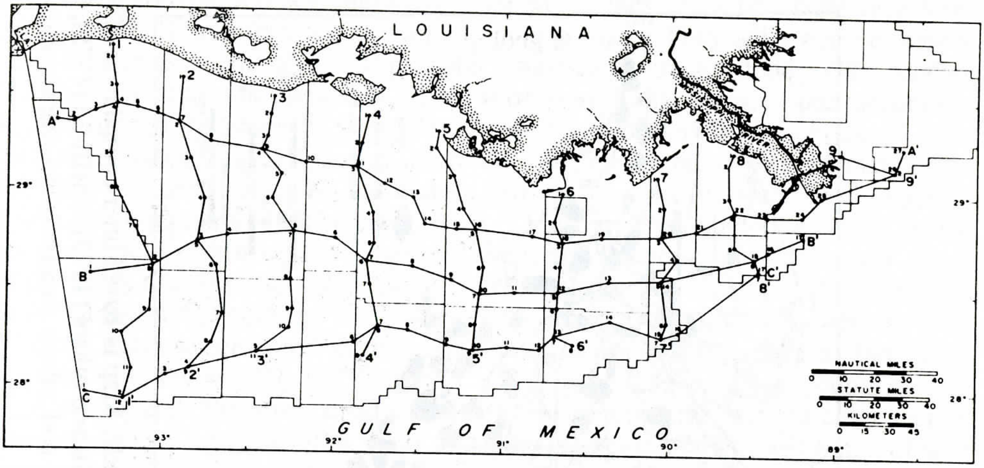


Figure 3. Locations of the cross-sectional network and individual well sites used in the study (from Shideler, 1986). The network consists of nine dip sections (1-9) and three strike sections (A-C); individual well sites indicated by small numerals.

of diagnostic benthonic and planktonic foraminifera and nannoplankton. The chronostratigraphic units of the studied interval range from early Miocene (Burdigalian) to late Pleistocene (Wisconsinan) in age. The locations of major faults and salt-dome structures were obtained mainly from a compilation of major structural features reported in the literature for the Louisiana shelf. In addition to reported faults, some unreported faults were inferred to be present along the cross-sectional lines on the basis of local structural characteristics. The lithology of the studied stratigraphic interval was determined from induction-electrical well logs, using the spontaneous potential and amplified short normal curves to establish sand-shale proportions. The entire regional cross-sectional network and associated well logs are illustrated in detail elsewhere (Shideler, 1986), but some selected generalized cross-section examples are presented in this summary report.

RESULTS AND DISCUSSION

Structural Framework

Regionally, the structural framework of the studied Neogene-Quaternary section composing the Louisiana continental shelf consists of a gulfward dipping and thickening assemblage of terrigenous strata. Superimposed on this regional framework is a complex fabric of local deformational features. These local structures have resulted largely from intrabasinal geostratic stresses caused by pronounced sediment-loading effects during the accretionary progradation of the northern Gulf continental margin.

A compilation of the distribution of major local structures reported in the literature illustrates a complex network of structures throughout the Louisiana shelf (Fig. 4). Diapiric salt domes are abundant. They are structures that have resulted from the lateral flowage and vertical intrusion of an initially stratiform Jurassic salt deposit (Louann Salt), which was stressed and mobilized by

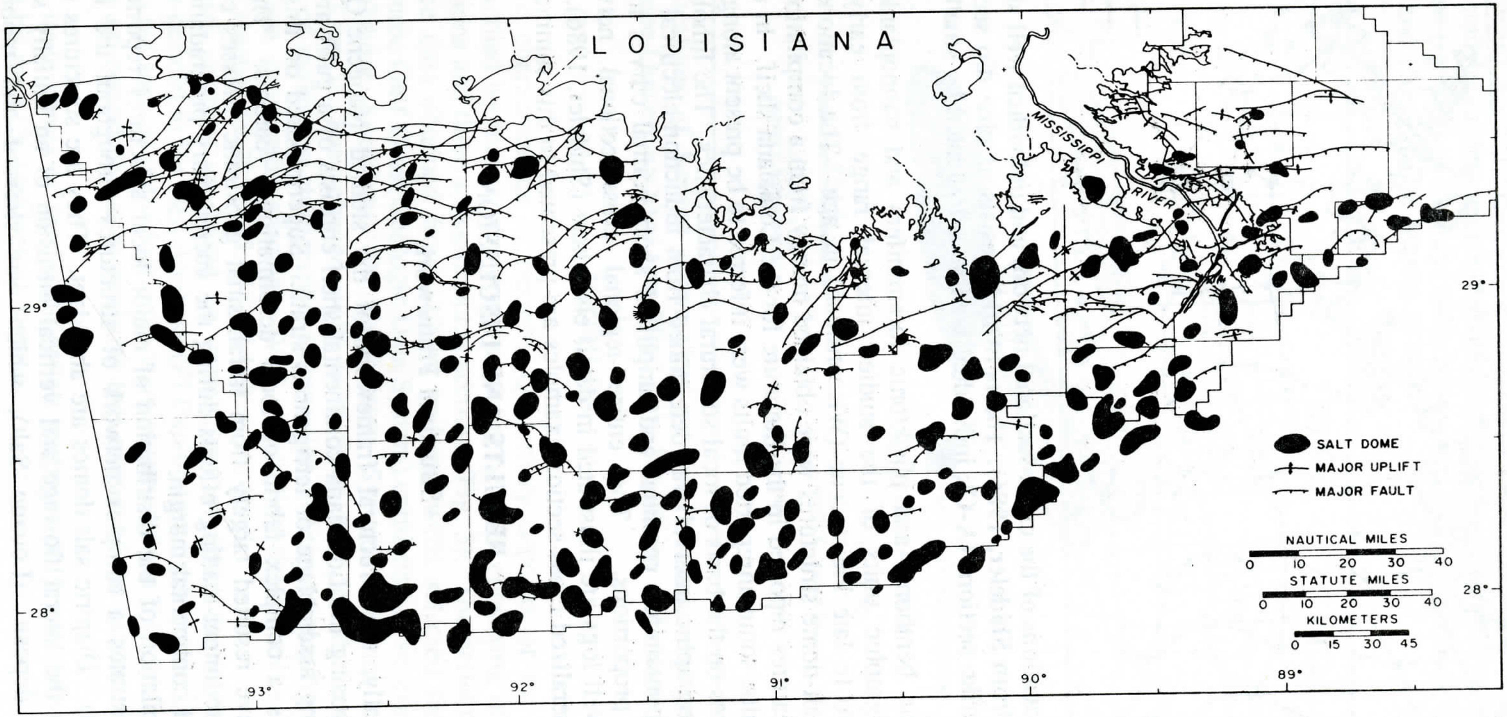


Figure 4. Distribution map of major structural features on the Louisiana continental shelf. Map is based on a compilation of major structures reported in the literature, mainly from the following sources: Braunstein and others (1973), Sheffield (1978), Martin (1980), U.S. Dept. of Interior (1982), Berryhill and Owen (1984), Berryhill and others (unpublished maps), and Chenoweth and McBride (1984).

differential overburden accumulation. The salt domes are a pervasive feature of the entire shelf, and they are of variable morphology. They appear to be mainly isolated diapirs, many of which probably coalesce at greater depth. The relative sizes of individual salt domes, in terms of cross-sectional areas, cannot accurately be determined because the compilation of reported domes is not based on a constant depth datum. However, there does appear to be an overall increase in both the size and density of salt domes gulfward toward the shelf break, such as noted by Woodbury and others (1973). Most of the diapirs, as well as the nonpiercement uplifts on the Louisiana shelf, are salt structures. However, some of the uplift features, especially along the outer shelf, may also represent mobilized shale intrusions (Woodbury and others, 1973; Martin, 1978, 1980). The network of cross sections prepared during the present study traverse several individual diapiric salt domes (Fig. 3), some of which penetrate strata as young as late Pleistocene age. However, most of the salt domes along the lines of cross section are deep-seated structures that occur at depths greater than the lower limits of well penetration.

Other common structural features of the Louisiana shelf are systems of large-scale growth faults oriented parallel or subparallel to depositional strike, generally in an east-west direction. The growth faults formed contemporaneously with deposition, largely in response to differential sediment loading and gravity-failure along former shelf-edge flexures. These gravity faults are characterized by the following: a down-to-basin displacement; a notable thickening of displaced strata on the gulfward downthrown side relative to time-equivalent strata on the upthrown side; and a progressive increase in stratigraphic throw with increasing depth. Growth faults on the Louisiana shelf tend to be progressively younger in a basinward direction (Martin, 1978), thus indicating the continued progradation of an accretionary shelf edge during Cenozoic time.

In addition to syndepositional growth faults, systems of post-depositional gravity faults are also common features on the Louisiana shelf. These are mainly tensional faults, and many are associated with salt dome structures where they have developed over the crests or along the flanks of individual domes. These gravity faults occur individually, in horst or graben structures, or in systems that form parallel or radial patterns.

The compilation of reported structural features on the Louisiana shelf (Fig. 4) is intended as a generalized representation illustrating the distribution of major structures. With few exceptions, the compilation probably accurately denotes the distribution of most of the known salt-dome/shale-intrusion structures on the shelf. However, the illustrated distribution of fault structures is probably less complete. Thousands of additional faults are present on the Louisiana shelf; however, available nonproprietary data are not sufficient to accurately delineate all of the numerous individual faults. The apparent gulfward reduction in the abundance of faults is probably an artifact resulting mainly from the relative scarcity of reported subsurface data from the deeper outer shelf region.

Stratigraphic Framework

The stratigraphic interval penetrated by wells of the cross-sectional network ranges from early Miocene to late Pleistocene in age. Similar to other stratigraphic intervals of the northern Gulf Basin, the studied interval illustrates a general overall thickening gulfward; much of the thickening occurs across systems of down-to-basin growth faults. The following discussion is a summary of the general characteristics of the studied interval, which is illustrated by means of selected representative dip and strike cross sections (Figs. 3, 5, 6); refer to Figure 1 for geographic areas. All chronostratigraphic units discussed are in accordance with those presented in the PI Exploration Systems Division (1985) database. The biohorizons (foraminifera and nannoplankton extinction horizons) establishing the boundaries of chronastratigraphic units in the database are based on biostratigraphic data provided by Paleo-Data, Inc.

Miocene section.— The studied Miocene section of the Louisiana shelf is

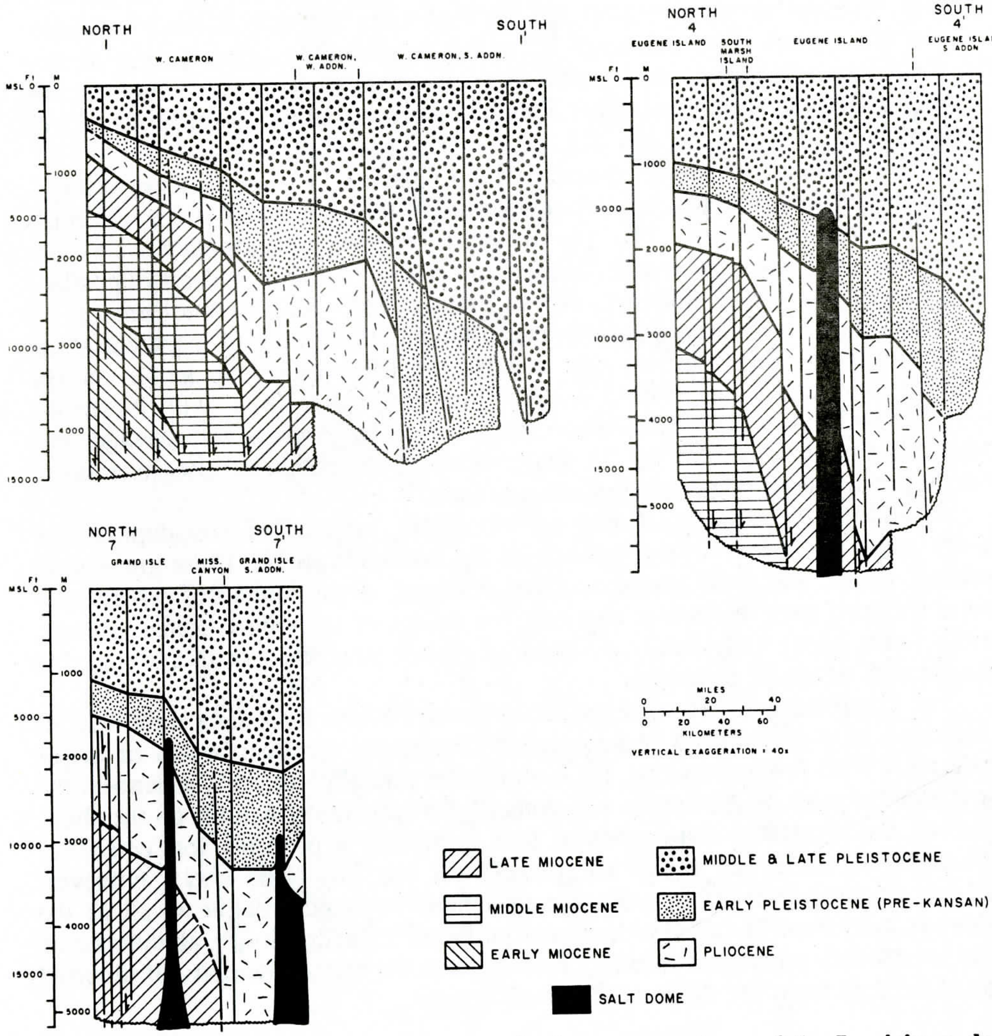


Figure 5. Representative dip-oriented geologic cross sections of the Louisiana shelf illustrating regional structure and stratigraphy of the western (1-1'), central (4-4'), and eastern (7-7') shelf sectors (see Fig. 3 for locations). Cross sections are generalized from Shideler (1986).

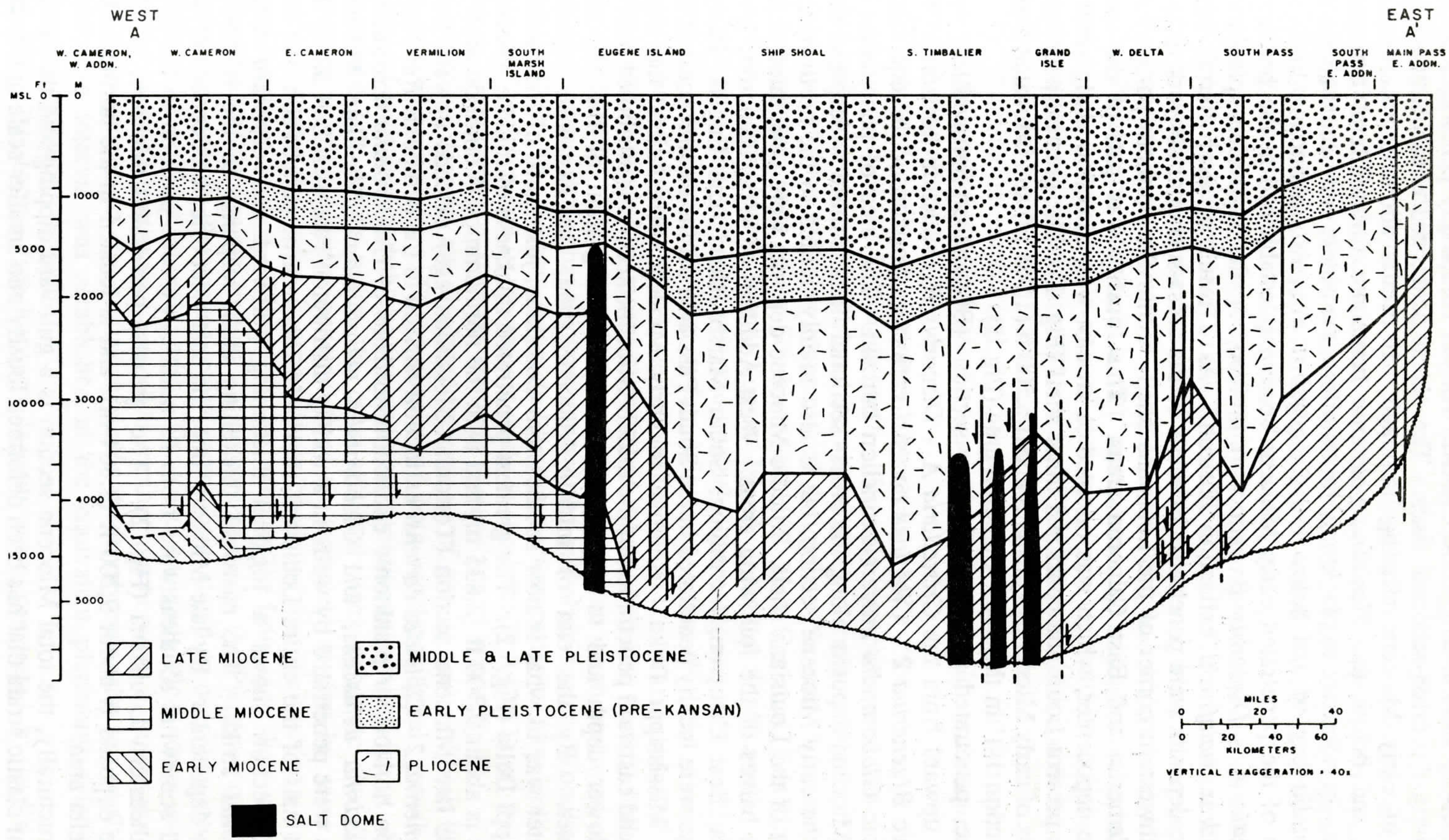


Figure 6. Representative strike-oriented geologic cross section (A-A') of the Louisiana shelf illustrating regional structure and stratigraphy of the inner shelf sector (see Fig. 3 for location). Cross section generalized from Shideler (1986).

represented by the Fleming Group, which was not completely penetrated by any wells along the cross-sectional lines. The oldest penetrated chronostratigraphic unit is of early Miocene (Burdigalian) age. It occurs above the *Lenticulina hanseni* and below the *Planulina palmerae* benthonic foraminiferal extinction horizons. In reference to planktonic foraminifera, the oldest unit occurs above the *Globorotalia kugleri* and below the *Catapsydrax dissimilis* extinction horizons. The top of the penetrated early Miocene section is established at the benthonic *Cristellaria A - Discorbis B* extinction horizon, or the equivalent planktonic *Catapsydrax stainforthi* extinction horizon. As a result of deep burial, early Miocene deposits were penetrated by wells of the cross-sectional network only in the northwestern corner of the Louisiana shelf, within the northern parts of the West Cameron and East Cameron areas. This locality was part of the early Miocene depocenter, which was oriented in an east-west direction along what is now the present Louisiana coast and inner shelf (Fig. 2). The greatest penetrated thickness of early Miocene deposits is about 6,500 ft (1,981 m) and is located along cross section 1-1' in the West Cameron area (Fig. 5).

The penetrated middle Miocene section (Burdigalian-Serravallian age) extends upward from the *Cristellaria A - Discorbis B* extinction horizon to the benthonic *Bigenerina 2 - Cibicides cartensi* extinction horizon or the equivalent planktonic *Globorotalia mayeri* extinction horizon. The area of penetration of middle Miocene deposits along the cross- sectional lines is more widespread than that of the early Miocene section, but is also mainly limited to the northwestern quadrant of the Louisiana shelf. Middle Miocene deposits were penetrated in the northern halves of the following areas: West Addition of West Cameron, West Cameron, East Cameron, Vermilion, South Marsh Island, and Eugene Island. They also were locally penetrated in the South Pass area off the eastern coast of the modern Mississippi Delta. Relative to early Miocene penetration, the more widespread eastward penetration of the middle Miocene section is attributed both to its shallower depth and to a progressive eastward migration of the Miocene depocenter. By the end of middle Miocene time, the east-west oriented depocenter was in what is now southeastern Louisiana, just north of the modern Mississippi Delta (Fig. 2). The greatest penetrated thickness of middle Miocene deposits is about 9,300 ft (2,835 m) and is in the northernmost Vermilion area.

The late Miocene section (Tortonian-Messinian age) extends upward from the *Bigenerina 2 - Cibicides cartensi* extinction horizon to the benthonic *Robulus E* extinction horizon, or planktonic extinction horizons of *Globorotalia acostaensis*, *Globoquadrina dehiscens*, and *Globorotalia merotumida*. The late Miocene deposits were penetrated by wells in a widely distributed east-west belt along the northern half of the entire Louisiana shelf. In the area of penetration, the late Miocene section shows a regional thickening gulfward and eastward along depositional strike. The eastward thickening is in response to the migrating Miocene depocenter. By late Miocene time, the east-west oriented depocenter had migrated somewhat southeastward from its middle Miocene position to the shelf off southeastern Louisiana (Fig. 2). The greatest penetrated thickness of late Miocene deposits is about 9,200 ft (2,804 m), and is located in the northernmost West Delta area.

Structurally, the total Miocene section is a gulfward dipping and thickening wedge of clastic strata that has been deformed locally into smaller-scale uplifts and basins, largely by deep-seated diapiric activity. Along the lines of cross-section, all observed diapirs are piercement salt domes that completely penetrate the Miocene section, illustrating that diapiric emplacement continued into post-Miocene time. The Miocene section is also extensively displaced by systems of gravity faults of both Miocene and post-Miocene age. These fault systems consist of both syndepositional growth faults, across which much of the sectional thickening occurs, and post-depositional tension faults.

Pliocene section.—The lithostratigraphic unit representing the total Pliocene section of the Louisiana shelf is the Citronelle Group. The base of the Pliocene section is established at the *Robulus E* extinction horizon, whereas the top of the Pliocene section is controversial. As noted by Poag and Valentine (1976), the Pliocene-Pleistocene boundary is not represented by a single biohorizon, but occurs within a stratigraphic interval that contains several different biohorizons. For the purpose of the present study, the top of the Pliocene section is defined at the benthonic *Buliminella 1* extinction horizon where present. In its absence, the top of the Pliocene is established at the extinction horizons of the planktonic *Globigerina nepenthes*, *Globorotalia margaritae*, *Globorotalia multicamerata*, or *Globoquadrina altispira*.

Pliocene deposits are penetrated by wells of the cross-sectional network throughout most of the Louisiana shelf. The only exception is within a deeply buried east-west oriented belt along the western half of the outer shelf. This belt was an area of thick sediment accumulation during late Pliocene-Pleistocene time, and encompasses parts of the South Additions of West Cameron, East Cameron, Vermilion, South Marsh Island, Eugene Island, and Ship Shoal areas. During Pliocene time, the east-west oriented basinal depocenter migrated southwesterly across the Louisiana shelf from its late Miocene location (Fig. 2). During early Pliocene time (3.5-6 m.y. B.P.), the depocenter was in the central mid-shelf sector, and by late Pliocene-Pleistocene time it had migrated to the western half of the outer-shelf sector. This migration resulted in deposition of an east-west oriented lens of Pliocene strata that thickens toward the mid-shelf sector (Woodbury and others, 1973). The thickest complete Pliocene section penetrated by wells of the cross-sectional network in the present study is about 8,600 ft (2,621 m) thick, and is along cross section 4-4' in the southern Eugene Island area (Fig. 5).

Pliocene deposits are displaced by systems of gravity faults of both Pliocene and post-Pliocene age. These consist of both syndepositional growth faults associated with sectional thickening and post-depositional tension faults. Several piercement salt domes along the cross-sectional lines penetrate the Pliocene section to varying degrees; penetration of the complete section by some of the domes illustrates that some diapiric activity continued into Pleistocene time.

Quaternary section.—Pleistocene deposits on the Louisiana shelf range from Nebraskan to Wisconsinan in age. In the present study, the Pleistocene section is divided into an early Pleistocene (pre-Kansan) interval and an overlying interval of middle and late Pleistocene deposits. The top of the early Pleistocene interval is established at the benthonic *Lenticulina 1* extinction horizon, a biohorizon that is widely recognized on the Louisiana shelf. In its absence, the top of the early Pleistocene interval was established at the equivalent planktonic *Globorotalia miocenica* extinction horizon.

Early Pleistocene (pre-Kansan) deposits were penetrated by wells of the cross-sectional network throughout most of the Louisiana shelf. Similar to Pliocene deposits, the only exception is within an east-west oriented belt along the western half of the outer shelf, which contains a relatively thick Pleistocene section. This region was part of the basinal depocenter throughout late Pliocene-Pleistocene time (Fig. 2). Consequently, deep burial precluded the penetration of early Pleistocene deposits in some parts of the depocenter region. The thickest observed section of early Pleistocene deposits penetrated by cross-sectional wells is more than 6,400 ft (1950 m) thick, and is in the southern East Cameron, South Addition area.

Middle and late Pleistocene (Kansan and post-Kansan) deposits are distributed throughout the Louisiana shelf. Greatest thicknesses are within the late Pliocene-Pleistocene depocenter along the western outer shelf. Locally, the greatest observed thickness of middle and late Pleistocene deposits penetrated by wells of the cross-sectional network is more than 12,000 ft (3,658 m), and is along cross section 1-1' in the southern part of the West Cameron, South Addition area (Fig. 5). During the last glacio-eustatic lowering of sea level (late Wisconsinan), the depocenter area contained a shelf margin, fluvial-deltaic complex of the ancestral Mississippi River, and sedimentation was highly controlled by active salt diapiric movements that influenced river-discharge patterns (Suter and Berryhill, 1985). Similar diapirically influenced shelf-margin delta complexes may account for most of the progradation of the Louisiana shelf during Neogene-Quaternary time.

The total Pleistocene section is a thick lenticular body of sediments with an east-west elongation along depositional strike (Woodbury and others, 1973). Structurally, Pleistocene deposits along the lines of cross section have been displaced by several systems of both syndepositional growth faults and post-depositional tension faults. Several piercement salt domes also penetrate the Pleistocene section. Most of the observed salt domes penetrate only early Pleistocene (pre-Kansan) deposits; however, a few domes do penetrate the middle-late Pleistocene section.

Holocene deposits were not differentiated from the Pleistocene section during this study because of the absence of shallow (<500 ft depth) subsurface well data. The surface casings of cross-sectional wells were usually set at depths below the shallow Holocene section; as a result, well samples and electric logs generally were not obtained for this near-surface interval. However, previous studies of this shallow section based on high-resolution seismic surveys of the Louisiana shelf (Berryhill, 1982; Berryhill and Suter, 1984; Suter and Berryhill, 1985) indicated that Holocene deposits show a regional variation in thickness; maximum thickness is about 500 ft (152 m) near the modern Mississippi Delta, and decreases toward the western end of the shelf where the deposits are either very thin (<10 ft or 3 m) or absent. In addition to delineating the uneven distribution of Holocene shelf deposits, the aforementioned studies also indicated the presence of lithofacies variations and the presence of young faults that offset parts of the Holocene section.

Regional Lithofacies Relationships

It has long been recognized that the deposition of successively younger offlapping wedges of siliciclastic sediment in the northern Gulf Basin during Cenozoic time resulted in the development of three regional, coast-parallel magnafacies. These lithofacies are genetically related to major deltaic depositional systems that transported the voluminous quantities of sediment into the Gulf Basin (Limes and Stipe, 1959; Thorsen, 1964; Norwood and Holland, 1974; Caughey, 1975). The magnafacies are based on sand-shale proportions, reflecting variations in depositional environments. In a gulfward direction, downdip facies transitions consist of a *massive sand facies* (>40 percent sandstone), an *alternating sand-shale facies* (15-40 percent sandstone), and a *massive shale facies* (<15 percent sandstone).

In the present study, the three magnafacies were found to be developed throughout most of the studied section, as determined from induction-electrical well logs. Regional lithofacies patterns within the early Miocene section were barely discernible because of the extremely limited area in which the section was penetrated; where penetrated, early Miocene deposits were of the massive sand facies and the alternating sand-shale facies. The area of penetration for middle Miocene deposits also was too limited to clearly delineate regional lithofacies patterns. However, where penetrated, dip-oriented cross sections did illustrate a gulfward transition from massive sand, to alternating sand-shale, to massive shale facies within the middle Miocene section. The area of penetration for the remainder of the studied section was sufficient to permit the delineation of well-defined regional lithofacies patterns. Deposits ranging in age from late Miocene to late Pleistocene show downdip transitions from massive sand, to alternating sand-shale, to massive shale magnafacies in a gulfward direction (Fig. 7).

In relating magnafacies distribution to the paleogeography of the basin's deltaic depositional systems, I used a lithofacies mapping method proposed by Norwood and Holland (1974), which is based largely on a facies model presented by Fisher (1969) for delta systems of the Gulf Basin. In this method, the *massive sand facies* (>40 percent sandstone) is interpreted to be representative of the following environments: fluvial channels, delta plain (distributary channel, levee, interdistributary, crevasse), distributary mouth bars, and marginal delta front environments. The *alternating sand-shale facies* (15-40 percent sandstone) is representative of neritic continental shelf or delta front (delta front slope, distal delta front) environments. The *massive shale facies* (<15 percent sandstone) represents a bathyal continental slope or prodelta slope environment. Norwood and Holland (1974) noted that the regional mapping of these three magnafacies within a given chronostratigraphic interval can provide an indication of the general framework of multiple delta systems within the selected interval. Of particular significance is the distribution of the massive sand facies, which generally outlines the morphology of composite deltaic progradations and shoreline positions during the mapped time interval.

The regional distribution of the three magnafacies delineated in the present study illustrates substantial temporal variability during Neogene-Quaternary time (Fig. 7). One variable aspect is the generally progressive gulfward displacement of the shelf lithofacies pattern. In late Miocene time, distribution of the massive sand facies was restricted to the present coastal and inner-shelf sector. During Pliocene time, the massive sand facies had expanded gulfward to cover approximately the northern third of the present shelf. Gulfward expansion of the massive

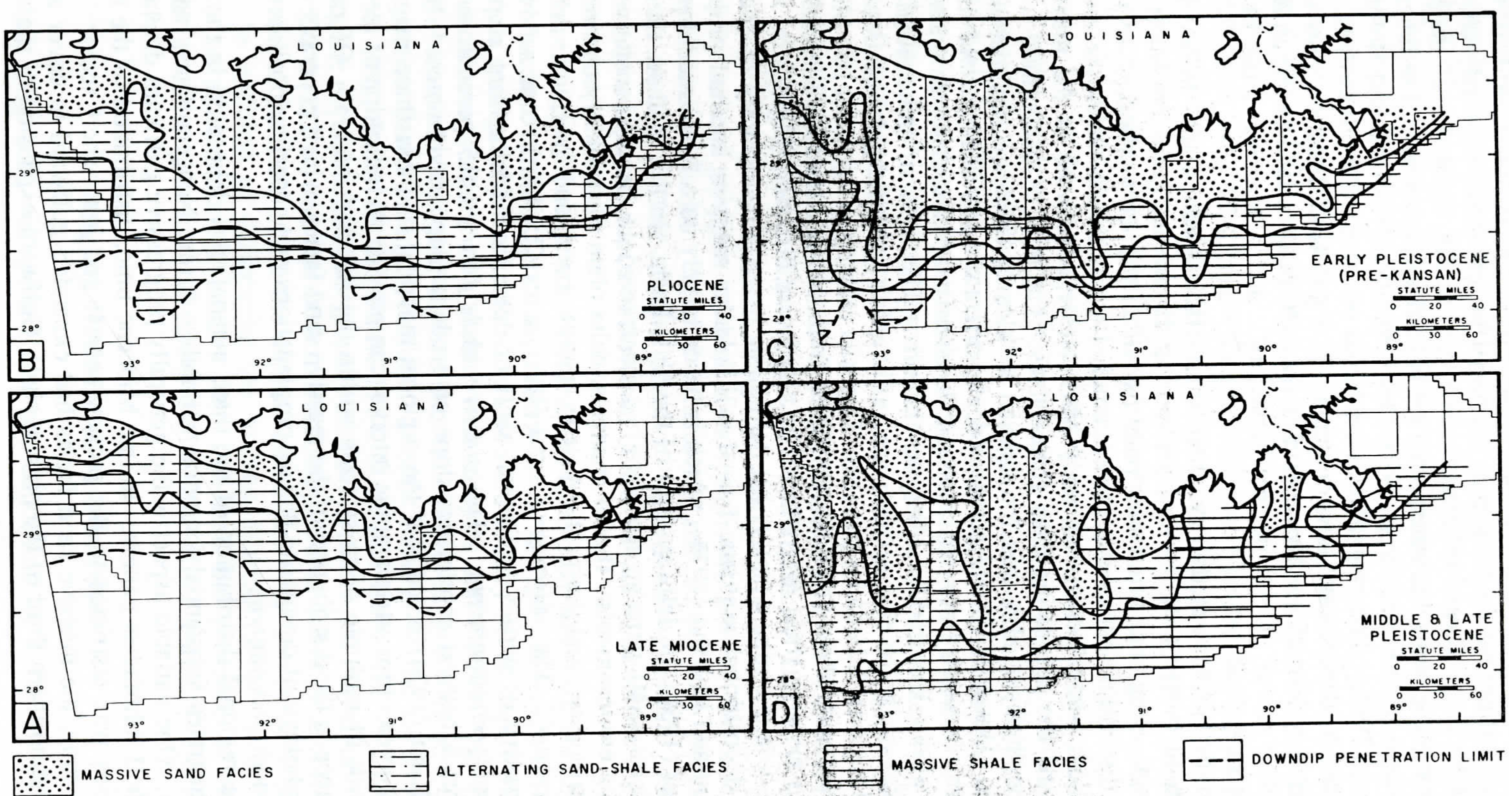


Figure 7. Lithofacies maps illustrating the regional distribution and temporal variability of three coast-parallel magnafacies based on sand-shale proportions. Facies are based on the mean composite lithologic characteristics of the late Miocene (A), Pliocene (B), early Pleistocene (C), and middle-late Pleistocene (D) stratigraphic sections.

sand facies continued during Pleistocene time, until it covered approximately the northern half of the present shelf. This gulfward displacement of the magnafacies transitions with time reflects a regressive sequence developed during the southward accretionary progradation of the Louisiana shelf margin.

A second temporally variable aspect of the regional magnafacies pattern is the changing gross morphology of the massive sand facies, which generally outlines composite deltaic progradations and approximate shoreline positions. Using the delta facies models presented by Fisher (1969), I interpret the basinal delta systems during Neogene-Quaternary time as being mainly of the lobate and elongate high-constructive varieties. These delta systems are characterized by high-volume sediment imput and component facies that are mainly influenced by fluvial processes, as opposed to marine processes. During late Miocene and Pliocene time, the deltaic progradations were mainly of the lobate variety (Figs. 7A,B). However, during early Pleistocene time, the deltas had become distinctly more elongated in morphology (Fig. 7C). By middle and late Pleistocene time, the deltas had become highly elongated (Fig. 7D). This temporal change in the gross morphology of deltaic progradations could reflect evolutionary changes of the deltaic systems during late Cenozoic time. As noted by Galloway (1975), morphological changes of a delta system can reflect progressive changes in the relative intensity of the various deltation processes; these process changes, in turn, may result from changes in the morphology of the receiving basin during delta development. One plausible explanation for the delta morphologic changes observed in the present study is a progressive transition from mainly wave-modified, fluvial-dominated lobate deltas in Miocene-Pliocene time, to mainly fluvial-dominated elongate deltas with reduced wave influence during Pleistocene time. This transition could have resulted from various possible causes, such as the following: (1) a progressive decrease in coastal wave energy flux for reworking deltaic margins, as a result of increased wave energy dissipation along the seafloor of an expanding shelf; (2) a relative increase in sediment influx during the Pleistocene, resulting in a greater emphasis of the constructive and net regressive deltation processes of aggradation and progradation, as opposed to destructive marine reworking processes.

A second possible contributing factor to the observed gross delta morphologic changes could be changes in load characteristics of the delta systems during late Cenozoic time. As noted by Fisher (1969), the elongate type of delta forms where the sediment load has a relatively high mud content and the deltaic sand bodies prograde over a relatively thick prodelta mud sequence. Upon delta abandonment, rapid differential subsidence of the sand facies results in permanent storage of most of the prograded elongate sand bodies. In contrast, lobate deltas form where the mud load content is relatively low and the sand bodies prograde over a relatively thin mud sequence. Under these conditions, differential subsidence of the sand bodies during abandonment is less rapid, and a substantial amount of the sand facies is reworked by marine processes into a broader lobate pattern. By this rationale, the morphologic changes of deltaic progradations noted in the present study could reflect a progressive increase in the relative mud content of the sediment load supplied by the basin's delta systems since late Miocene time. This might be attributed to a combination of factors. Certainly, the onset of Pleistocene continental glaciation in North America may have resulted in substanflowing drainage systems. In addition, as gulfward progradation of the Louisiana shelf margin continued, the basin's delta systems became progressively more remote from Rocky Mountain and mid-continent sediment source areas. This factor, in conjunction with a progressive denudation of the source areas, also may have contributed to increasingly greater proportions of argillaceous sediments deposited within the northern Gulf of Mexico Basin. In turn, increased argillaceous influx may have contributed to the morphologic changes of the basinal delta systems during late Cenozoic time.

ACKNOWLEDGMENTS

The author expresses his appreciation to R.H. Wells and L.M. Massingill, who provided valuable assistance during various phases of the study. Appreciation is also expressed to R.Q. Foote for many beneficial discussions during the course of the study, and to D.J.P. Swift, M. Ball and J.R. Suter for reviewing the manuscript.

REFERENCES CITED

Berryhill, H.L., Jr., 1982, Ancient fluvial systems and shorelines on the continental shelf off southwest Louisiana, *in* Nummedal, D., ed., Sedimentary processes and environments along the Louisiana-Texas coast: Geological Society of America, Field Trip Guidebook, p. 87-91.

Berryhill, H.L., Jr., and Owen, D.E., 1984, Shallow geology of the continental slope: Location of diapris, fold axes, and faults: Minerals Management Service OCS Map Series, MMS 84-0003, sheet 3.

Berryhill, H.L., Jr., and Suter, J.R., 1984, Continental shelf sands, northwest Gulf of Mexico (abs.): Sedimentology of Shelf Sands and Sandstones Research Symposium Proceedings, Canadian Society of Petroleum Geologists, Calgary, p. 25.

Braunstein, J., Hartman, J.A., Kane, B.L., and van Amringe, J.H., eds., 1973, Offshore Louisiana oil and gas field: New Orleans and Lafayette Geological Societies, 124 p.

Caughey, C.A., 1975, Pleistocene depositional trends host valuable Gulf oil reserves: Oil and Gas Journal, v. 73, nos. 36, 37, p. 90-94, 240-242.

Chenoweth, P.A., and McBride, M.H., 1984, Principal Structural Features of Louisiana Map, Pennwell Publishing Co., Tulsa.

Coleman, J.M., Prior, D.B., and Lindsay, J.F., 1982, Formation of the Mississippi Canyon (abs.): Gulf Coast Association of Geological Societies Transactions, v. 32, p. 519.

Curtis, D.M., and Picou, E.B., Jr., 1978, Gulf Coast Cenozoic: a model for the application of stratigraphic concepts to exploration on passive margins: Gulf Coast Association of Geological Societies Transactions, v. 28, p. 103-120.

Fisher, W.L., 1969, Facies characterization of Gulf Coast Basin delta systems, with some Holocene analogues: Gulf Coast Association of Geological Societies Transactions, v. 19, p. 239-261.

Galloway, W.E., 1975, Process framework for describing the morphologic and stratigraphic evolution of deltaic depositional systems, *in* Broussard, M.L.,

ed., Deltas - Models for Exploration: Houston Geological Society, p. 87-98.
Hardin, G.C., Jr., 1962, Notes on Cenozoic sedimentation in the Gulf Coast geosyncline, U.S.A., *in* Geology of the Gulf coast and Central Texas and Guidebook of Excursions: Geological Society of America Annual Meeting and Houston Geological Society, p. 1-15.
Limes, L.L., and Stipe, J.C., 1959, Occurrence of Miocene oil in south Louisiana: Gulf Coast Association of Geological Societies Transactions, v. 9, p. 77-90.
Lynch, C.W., and Rudolph, R.W., 1984, Gulf of Mexico Summary Report - Outer Continental Shelf Oil and Gas Activities in the Gulf of Mexico and their Onshore Impacts: U.S. Dept. of Interior, Minerals Management Service, OCS Information Report MMS 84-0074, 108 p.
Martin, R.G., 1978, Northern and eastern Gulf of Mexico continental margin: stratigraphic and structural framework, *in* Bouma, A.H., Moore, G.T., and Coleman, J.M., eds., Framework, Facies, and Oil-trapping Characteristics of the Upper Continental Margin: AAPG Studies in Geology No. 7, p. 21-42.
Martin, R.G., 1980, Distribution of salt structures in the Gulf of Mexico: U.S. Geological Survey Misc. Field Studies Map MF-1213.
Murray, G.E., 1961, Geology of the Atlantic and Gulf Coastal Province of North America: Harper and Bros., New York, 692 p.
Norwood, E.M., Jr., and Holland, D.S., 1974, Lithofacies mapping: a descriptive tool for ancient delta systems of the Louisiana Outer Continental Shelf: Gulf Coast Association of Geological Societies Transactions, v. 24, p. 175-188.
PI Exploration Systems Division, 1985, Gulf Coast Cenozoic Stratigraphic Chart, Petroleum Information Corp., Houston.
Pilger, R.H., ed., 1980, The origin of the Gulf of Mexico and the early opening of the central North Atlantic Ocean: Louisiana State University Dept. of Geology and Louisiana Geological Survey Symposium Proceedings, Baton Rouge, 98 p.
Poag, C.W., and Valentine, P.C., 1976, Biostratigraphy and ecostratigraphy of the Pleistocene basin, Texas- Louisiana continental shelf: Gulf Coast Association of Geological Societies Transactions, v. 26, p. 185-226.
Rainwater, E.H., 1967, Resume of Jurassic to recent sedimentation history of the Gulf of Mexico Basin: Gulf Coast Association of Geological Societies Transactions, v. 17, p. 179-210.
Sheffield, F.C., 1978, Where to next in the Gulf of Mexico? A brief review of future exploration opportunities in the Gulf: Offshore Technology Conference Proceedings, Houston, v. 1, p. 383-387.
Shideler, G.L., 1986, Regional geologic cross-section series of Neogene-Quaternary deposits, Louisiana continental shelf: Geological Society of America Map and Chart Series, MC-54.
Suter, J.R., and Berryhill, H.L., Jr., 1985, Late Quaternary shelf-margin deltas, northwest Gulf of Mexico: AAPG Bulletin, v. 69, p. 77-91.
Thorsen, C.E., 1964, Miocene lithofacies in southeast Louisiana: Gulf Coast Association of Geological Societies Transactions, v. 14, p. 193-201.
U.S. Dept. of Interior, 1982, Gulf of Mexico OCS Environmental Map, Minerals Management Service OCS Regional Office, New Orleans, LA.
Woodbury, H.O., Murray, I.B., Jr., Pickford, P.J., and Akers, W.H., 1973, Pliocene

and Pleistocene depocenters, outer continental shelf, Louisiana and Texas: AAPG Bulletin, v. 57, p. 2428-2437.

MAJOR CHEMICAL CHARACTERISTICS OF THE HAMMETT GROVE META-IGNEOUS SUITE, NORTHWESTERN SOUTH CAROLINA

STEVEN K. MITTWEDE — *South Carolina Geological Survey, Harbison Forest Road, Columbia, SC 29210*

MAGNE ØDEGÅRD — *Geological Survey of Norway, Chemistry Division, Post Box 3006 N-*7001 Trondheim, Norway

W.E. SHARP — *Department of Geology The University of South Carolina Columbia, SC 29208*

ABSTRACT

Twenty-five major-oxide analyses show that rocks of the Hammett Grove meta-igneous suite of northwestern South Carolina are comparable to those of well-documented ophiolite complexes, including those of the southern Appalachians. Based upon these analyses, the rocks of the suite are divisible into six major groups: altered ultramafites, metapyroxenite, high-CaO metapyroxenite, metagabbro, metaplagiogranite and metabasalt. Although the rocks of the suite have undergone amphibolite-grade metamorphism, all of the groups, except the high-CaO metapyroxenite, are distinguishable in the field, thereby facilitating mapping and sampling of the various lithologies.

Differentiation trends, defined on AFM and CAM ternary diagrams, and other geochemical relationships, are comparable to those of ophiolites described elsewhere. Moderate iron enrichment with respect to magnesium content in the metagabbros probably represents magmatic differentiation and a simple fractional crystallization model (olivine → clinopyroxene → calcic plagioclase) seems to viably explain the various metapyroxenite-to-metagabbro increases and decreases in major oxides. An AFM ternary plot and a plot of silica versus total iron/total iron + magnesia show that the Hammett Grove rocks, particularly the cumulate rocks of the suite, have generally narrow, overlapping compositional ranges, suggesting that the metapyroxenite, high-CaO metapyroxenite and metagabbro are comagmatic. The plot of oxide analyses on a CAM ternary diagram demonstrates that the cumulate rocks, though comagmatic and part of a roughly linear differentiation trend, are clearly divisible into distinct geochemical groups. Although mainly applied to basaltic lavas, an alkali-silica plot shows that the Hammett Grove metapyroxenites, metagabbro and metabasalt, all probably derived from a basaltic parent magma, have tholeiitic affinity.

INTRODUCTION

Mittwede (1986) was the first to recognize and describe the 11-kilometer long, mafic-ultramafic rock body that is exposed in the early Paleozoic or late

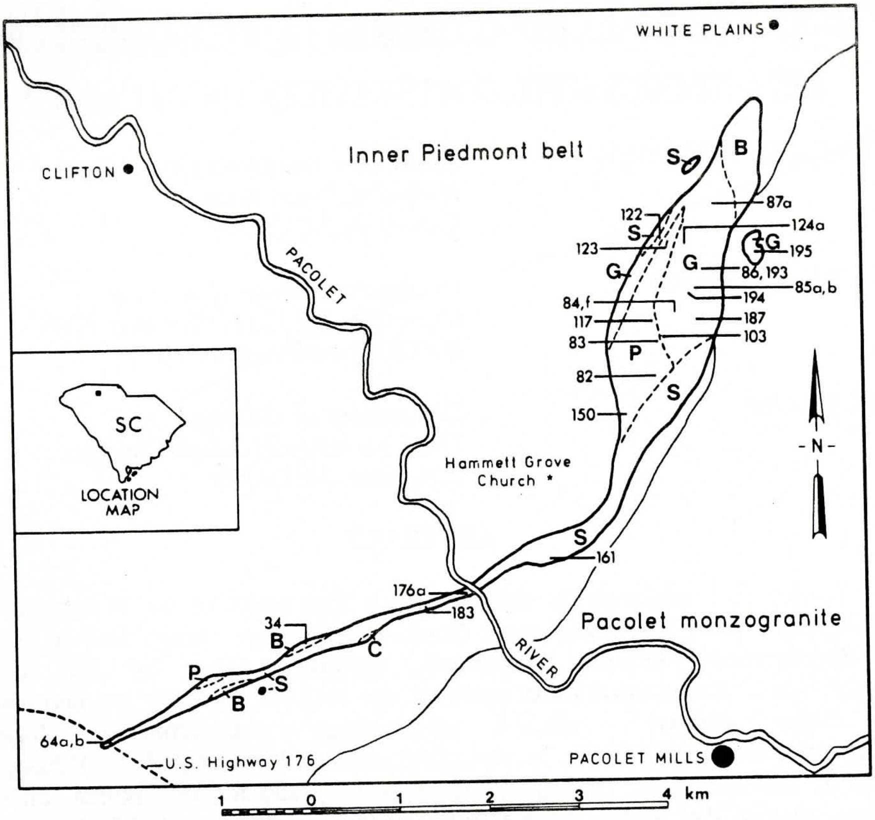


Figure 1. Map showing location of study area, distribution of rock types in the Hammett Grove meta-igneous suite and sample locations. Explanation: S - altered ultramafite, P - metapyroxenite, G - metagabbro, B - metabasalt, C - metachert.

Precambrian Inner Piedmont belt of eastern Spartanburg and southwestern Cherokee counties, South Carolina. This rock body, which is comprised mainly of amphibolite-grade metamorphosed igneous rocks, is here referred to informally as the Hammett Grove meta-igneous suite. Lieber (1858) and Sloan (1908) briefly described talcose ultramafite in the vicinity of Cedar Spring in eastern Spartanburg County, South Carolina, and the senior author now recognizes that ultramafite as one of a number of narrow, isolated bodies of ultramafic rock that occur along strike to the southwest of the main slab of the Hammett Grove suite, and that all are part of the same lithotectonic unit. The total strike-length of the suite, including the small rock bodies southwest of the main slab, is approximately 30 kilometers. Wagener and Price (1976) prepared a reconnaissance map of the Pacolet Mills 7.5-minute quadrangle and delineated a gabbro-amphibolite unit that corresponds, in part, to the northeasternmost portion of the Hammett Grove suite. None of the suite appeared on the small-scale map of the South Carolina Piedmont prepared by Overstreet and Bell (1965).

The Hammett Grove meta-igneous suite consists of distinctive mappable units including altered (steatitized and serpentinized) ultramafite, metapyroxenite, metagabbro (isotropic and locally layered) and amphibolitic metabasalt, the distributions of which are shown in Figure 1. Although contacts between the units are rarely exposed, there is no evidence that they are tectonic. In fact, the intimate field relationships of the rocks of the suite indicate that the contacts between units within the suite are gradational or locally intrusive. The complex map pattern of the units within the suite does not seem to support the possibility that the contacts between units might be faults.

The altered ultramafite unit consists of massive soapstone, impure talc schist and lesser serpentinite which locally contains relict olivine. The metapyroxenite and metagabbro are generally characterized by coarse to very coarse grain sizes, and most of these rocks, especially the metagabbros, display what is interpreted as relict cumulate texture. Microprobe-derived compositions (Table 1) of relict clinopyroxene (diopsidic augite) in the metapyroxenite and metagabbro and relict plagioclase (bytownite - An_{82} to An_{90}) in the metagabbro are similar to those of the ophiolitic Halifax County complex in the easternmost North Carolina Piedmont (Kite and Stoddard, 1984). Fine- to medium-grained felsic intrusive rocks, comprised mainly of quartz and plagioclase, are present within the metagabbro unit and are interpreted as metamorphosed plagiogranite (Coleman and Peterman,

Table 1. Representative analyses of primary minerals from the Hammett Grove meta-igneous suite.

	Clinopyroxene		Plagioclase	
	117A	122A	103C	122B
SiO_2	54.21	54.43	46.34	47.83
Al_2O_3	1.26	1.15	35.59	34.62
TiO_2	0.13	0.05	-	-
FeO*	4.52	6.51	0.06	0.02
MnO	0.23	0.31	-	-
MgO	16.55	15.13	-	-
CaO	24.82	24.33	18.22	17.31
Na_2O	0.12	0.33	1.28	1.89
K_2O	0.05	0.06	0.07	0.06
Total	101.89	102.30	101.56	101.73

	Numbers of ions on the basis of 6 oxygens				Numbers of ions on the basis of 8 oxygens			
	117A		122A		103C		122B	
Si	1.921	}1.975	1.971	}2.020	2.101	}4.003	2.158	}3.999
Al	0.054		0.049		1.902		1.841	
Ti	0.004		0.001		-		-	
Fe^{3+}	0.136		0.197		0.002		0.001	
Mg	0.891		0.817		-		-	
Mn	0.007	2.008	0.010	1.995	-	1.004	-	1.008
Ca	0.960		0.944		0.885		0.837	
Na	0.008		0.023		0.113		0.166	
K	0.002		0.003		0.004		0.004	
Total	3.983		4.015		5.007		5.007	

* total iron as FeO; - not analyzed

117A - average of 5 spots from relict clinopyroxene in high-CaO metapyroxenite

122A - average of 5 spots from relict clinopyroxene in very coarse-grained metagabbro

103C - average of 10 spots from relict plagioclase in medium- to coarse-grained metagabbro

122B - average of 10 spots from relict plagioclase in very coarse-grained metagabbro

1975). Fine-grained, dense, bedded quartzite, interpreted as metachert, occurs at one locality. Mittwede (1986) noted the comparability of this rock assemblage to the well-known Steinmann trinity and suggested that it might represent metamorphosed, dismembered ophiolite.

Preliminary observations of the geometry, structure and contact relationships (Mittwede, 1986) of the Hammett Grove suite indicate that it is a thin (≤30 meters vertically), horizontal to sub-horizontal, allochthonous thrust sheet. The basal contact of the suite, observed at only one locality, is a highly deformed shear zone along which massive serpentinite of the Hammett Grove suite and biotite gneiss of the Inner Piedmont are juxtaposed. Intrafolial shear crenulations indicate that the basal shear zone records dextral motion and therefore, westward movement of the thrust sheet relative to the rocks of the Inner Piedmont (Allen Dennis, personal communication, 1987). The Hammett Grove suite may actually be a klippe as no upper contacts or overlying rocks have been identified unequivocally. Similarity to nearby Kings Mountain belt ultramafites and associated rocks may imply that these and the Hammett Grove rocks have a common origin (Mittwede, 1987). Presently, an emplacement model involving oblique collision and obduction seems the most viable for explaining the distribution of rock types, gross structural features and previously described characteristics of the Kings Mountain shear zone (Horton, 1981).

CHEMICAL ANALYSIS

Sampling and Initial Preparation

A group of twenty-five rock samples (Figure 1) was selected on the basis of their freshness from among rocks collected during geologic mapping of the main (11-km long) body of the Hammett Grove meta-igneous suite. Localities were not sampled systematically, but an effort was made to collect samples representative of the suite and its component lithologies. The selected samples were carefully trimmed on a water-cooled diamond saw to remove weathered material and organic matter. The samples were later run through a jaw crusher, and homogenization was accomplished through grinding in an agate swing mill.

Final Preparation and Instrumentation

Analytical portions of 0.2 grams were weighed into 100 ml polyethylene bottles, 5 ml of minimum 40 percent HF was added to each, and the bottles were sealed immediately with polyethylene caps. After one hour, the samples were left overnight on a steam bath. After cooling, 50 ml of 4.5 percent boric acid was added to complex the free HF and the samples were placed again on the steam bath to dissolve any precipitated fluorides. The samples were then diluted so that the original samples had been diluted by a factor of 10 000 yielding a direct correspondence between µg/ml and % in the original samples (Ødegård, 1979).

The samples were analyzed with an inductively coupled argon plasma (Jarrel-Ash 975 ICAP AtomComp) simultaneous spectrometer for the standard rock-forming elements Si, Al, Fe, Ti, Mg, Ca, Na, K, Mn, and P using B as a reference element (Ødegård, 1979). The precision of the analyses, using the Pageland diabase as a reference material, is estimated to have a coefficient of variation (s/m) of 0.02 except for K, Ti and P which have considerably larger coefficients of variation (0.13, 0.22 and 0.18 respectively). This is not serious because the K_2O, TiO_2 and P_2O_5 in these rocks is quite low and there could be a significant sub-sampling error if these elements occur in rare grains.

RESULTS AND DISCUSSION

Except in certain cases involving metapyroxenite and melanocratic metagabbro, the different lithologies of the Hammett Grove meta-igneous suite are visually distinctive, and these chemical analyses (Tables 2-7) confirm that field-use rock names are appropriate descriptors. The only rock type not distinguishable in the field, and that is, therefore, solely a chemical designation, is high-CaO metapyroxenite.

The chemical analyses were grouped on the basis of physical and chemical similarity and were tabulated, including chemical-group means and standard deviations (σn-1) for the metapyroxenite, high-CaO metapyroxenite and metagabbro groups (Tables 3-5). Normative mineral compositions for the chemical-group means or single analyses (metaplagiogranite and metabasalt) were obtained using a personal computer-C.I.P.W. program on floppy disk by Mantei and Varner.

In this study, it was assumed that the metamorphism which affected the rocks of the Hammett Grove meta-igneous suite was isochemical, except in the case of the altered ultramafites (soapstone, serpentinite) in which we believe metamorphism was probably quasi isochemical or allochemical (Winkler, 1979). Although the rocks have undergone recrystallization, both relict, primary minerals and textures remain in most of the cumulate rocks of the suite, demonstrating that the high- grade metamorphic event which affected them did not obliterate all of the original aspects of the rocks. Winkler (1979) cited the work of E. Jäger who found that, even in high-grade gneisses, the rock system was closed to K and Ca, elements generally assumed to be "mobile". Winkler also cited the work of Ronov and others who showed that "regional metamorphism of sediments is isochemical, apart from H_2O and CO_2" (p. 16). The altered ultramafites have obviously undergone hydrous metamorphic transformations (steatitization, serpentinization or both), and therefore their alteration was undoubtedly quasi isochemical (that is, isochemical except for the highly volatile components H_2O and CO_2) or allochemical. Furthermore, if metamorphism had been allochemical (or metasomatic), addition of mobile elements, not depletion, would have accompanied the introduction of metamorphic fluids or gases (H_2O and CO_2), as was noted by Scotford and Williams (1983). It must also be understood that even mobile elements generally do not leave the rock system during metamorphism; Winkler (1979) wrote: "Transport of material is generally limited to distances similar to the size of newly formed crystals" (p. 16). If there was some movement of alkalis, for instance, relative enrichment may have occurred on a limited scale, but the very low potassium and sodium contents of these mafic and ultramafic rocks do not support such a conclusion. For these reasons, we are confident that the major-oxide analyses reported here represent very nearly the original rock compositions (excepting of course the altered ultramafites).

Altered Ultramafite

Five samples of the altered ultramafite were analyzed (Table 2). Most of the altered ultramafite is soapstone or talc schist (talc±clinoamphibole±opaques± chlorite±serpentine±orthoamphibole), but locally, is mainly massive serpentinite (antigorite ±chlorite±olivine±talc±orthoamphibole±opaques). The high MgO and low Al_2O_3 contents suggest that these rocks were originally olivine-rich, probably dunite or peridotite.

Table 2. Analyses of altered ultramafites.

Sample number	64A	64B	161	176A	183
SiO_2	53.48	50.48	45.36	42.50	57.97
Al_2O_3	1.68	3.85	7.04	2.26	2.21
TiO_2	0.01	0.14	0.86	0.05	0.05
Fe_2O_3 *	4.22	6.77	14.59	9.56	6.28
MnO	0.07	0.09	0.15	0.16	0.08
MgO	25.51	23.98	26.24	34.91	27.18
CaO	0.90	2.34	1.78	0.72	0.26
Na_2O	nd	0.06	0.01	nd	nd
K_2O	0.26	nd	nd	nd	0.15
P_2O_5	nd	nd	0.20	nd	nd
Total **	86.13	87.71	96.23	90.16	94.18

64A - vermiculite-talc schist
64B - amphibole-vermiculite-talc schist
161 - impure soapstone (partially altered metapyroxenite)
176A - chloritic serpentinite
183 - talc schist

* Total iron as Fe_2O_3 .
** Note: Low totals are due to unmeasured water content of the hydrous silicates talc, vermiculite, chlorite and serpentine.

Table 3. Analyses of metapyroxenite.

Sample number	82	83	85A	123	150	$\bar{x}$	$\sigma n-1$
SiO_2	43.93	41.87	46.27	43.35	41.94	43.47	1.80
Al_2O_3	7.42	8.03	12.25	13.51	11.81	10.60	2.71
TiO_2	0.38	0.20	0.20	0.23	0.21	0.24	0.08
Fe_2O_3 *	11.02	13.30	9.46	11.24	12.55	11.51	1.48
MnO	0.17	0.18	0.12	0.13	0.22	0.16	0.04
MgO	21.56	24.76	20.84	20.40	20.03	21.52	1.90
CaO	9.06	6.74	9.82	8.49	7.86	8.39	1.17
Na_2O	0.22	0.36	0.64	0.59	0.70	0.50	0.20
K_2O	nd	nd	nd	0.13	nd	0.03	0.06
P_2O_5	nd	nd	nd	nd	nd	-	-
Total	93.76	95.44	99.60	98.07	95.32	96.42	

* Total iron as Fe_2O_3.

Normative mineral composition for the average metapyroxenite (assumed $FeO:Fe_2O_3$= 3:1): Mag - 4.17, Ilm - 0.46, Ab - 4.23, An - 26.56, Or - 0.18, Di - 12.00, En - 17.25, Ol - 31.58.

Metapyroxenite

Five samples of metapyroxenite were analyzed (Table 3). The metapyroxenite is composed mainly of magnesio-hornblende with lesser amounts of tremolite-actinolite, relict clinopyroxene, chlorite and opaques. The chemical-group mean compares favorably to the average pyroxenite compositions reported by Nockolds (1954) and especially to the average pyroxenite composition reported by Kite and Stoddard (1984) from the Halifax County complex in the Eastern Slate Belt of North Carolina. Calculation of the normative mineralogy for the average metapyroxenite analysis shows that the original lithology may have contained both plagioclase and olivine, as well as both orthopyroxene and clinopyroxene (Table 3). Therefore, the precursor may have been a plagioclase- bearing, olivine pyroxenite or plagioclase lherzolite.

High-CaO Metapyroxenite

Three samples of this rock were analyzed (Table 4). This rock type, composed mainly of magnesio-hornblende and relict clinopyroxene, occurs in units mapped as both metapyroxenite and metagabbro; it is visually indistinguishable from these units. Field relationships, though obscure due to lack of exposure, suggest that the high-CaO metapyroxenite may occur as dikes or small pod-like bodies (a localized product of differentiation) within the cumulate rocks of the suite. Its distinctive composition and the relatively small standard deviation for the group mean confirm that the high-CaO metapyroxenite is indeed a valid, separate rock group. In addition to being higher in CaO than the other metapyroxenite, the high-CaO metapyroxenite also contains notably lower Al_2O_3, Fe_2O_3 and MgO and appreciably higher SiO_2 and TiO_2. A normative mineral composition (Table 4) shows that the precursor for the high-CaO metapyroxenite was grossly similar to that of the metapyroxenite, but was relatively pyroxene-rich and plagioclase and olivine poor.

Table 4. Analyses of high-CaO metapyroxenite.

Sample number	117	187	194	$\bar{x}$	$\sigma n-1$
SiO_2	49.13	47.07	50.17	48.79	1.82
Al_2O_3	6.96	7.00	5.38	6.45	0.92
TiO_2	0.61	0.41	0.58	0.53	0.11
Fe_2O_3 *	8.89	8.12	8.48	8.50	0.39
MnO	0.16	0.14	0.16	0.15	0.01
MgO	17.36	18.54	17.62	17.84	0.62
CaO	12.68	12.53	13.67	12.96	0.62
Na_2O	0.61	0.83	0.58	0.67	0.14
K_2O	0.21	0.23	0.10	0.18	0.07
P_2O_5	nd	nd	nd	-	-
Total	96.61	94.87	96.74	96.07	

* Total iron as Fe_2O_3.

Normative mineral composition for the average high-CaO metapyroxenite (assumed $FeO:Fe_2O_3$= 3:1): Mag - 3.08, Ilm - 1.01, Ab - 5.66, An - 14.04, Or - 1.06, Di - 39.87, En - 22.14, Ol - 9.20.

Metagabbro

Ten samples of metagabbro were analyzed (Table 5). As would be expected, the metagabbros are the most compositionally variable of the rock units, and this is a function of the varying mafic mineral-plagioclase ratio in the gabbroic precursor. The metagabbros are composed mainly of magnesio-hornblende and plagioclase with lesser amounts clinozoisite, quartz, actinolite, relict clinopyroxene, muscovite, sphene and opaques. Normative mineralogy suggests that the average metagabbro was originally olivine gabbro, but no relict olivine is present in the metagabbros. The relatively high SiO_2, Al_2O_3, CaO and alkali contents, as compared to the metapyroxenite, are due to the presence of plagioclase (50 percent normative plagioclase). These metagabbro compositions are similar to those reported from southern Appalachian ophiolites by Kite and Stoddard (1984) from the Halifax County complex and by Drake and Morgan (1981) from the Piney Branch Complex in northern Virginia.

Table 5. Analyses of metagabbro.

Sample number	84	85B	86	87A-1	87A-2	103
SiO_2	48.34	44.66	43.95	48.71	47.67	45.86
Al_2O_3	12.65	15.33	21.00	17.70	19.69	17.74
TiO_2	0.36	0.27	0.22	0.58	0.49	0.67
Fe_2O_3 *	6.99	6.74	5.10	7.88	6.91	6.81
MnO	0.13	0.11	0.08	0.14	0.11	0.11
MgO	13.49	12.85	9.76	10.26	9.94	9.74
CaO	15.85	13.67	14.20	13.98	14.26	14.99
Na_2O	0.55	0.76	0.77	1.04	0.69	0.76
K_2O	0.49	0.20	0.18	0.34	0.18	0.12
P_2O_5	nd	nd	nd	0.06	nd	nd
Total	98.85	94.59	95.26	100.69	99.94	96.80

Sample number	122	124A	193	195	$\bar{x}$	$\sigma n-1$
SiO_2	50.05	48.32	44.68	48.98	47.12	2.15
Al_2O_3	19.76	20.51	21.19	14.01	17.96	3.04
TiO_2	0.46	0.36	0.22	0.54	0.42	0.16
Fe_2O_3	6.15	5.20	5.75	6.65	6.42	0.87
MnO	0.12	0.09	0.08	0.12	0.11	0.02
MgO	7.51	8.73	9.66	10.59	10.25	1.77
CaO	16.29	17.93	15.36	16.72	15.33	1.37
Na_2O	1.20	0.58	0.79	0.13	0.73	0.29
K_2O	0.32	0.13	0.14	nd	0.21	0.14
P_2O_5	nd	nd	nd	nd	0.01	0.02
Total	101.86	101.85	97.87	97.74	98.56	

* Total iron as Fe_2O_3.

Normative mineral composition for the average metagabbro (with all iron assumed to be FeO): Ilm - 0.80, Ab - 6.17, An - 45.06, Or - 1.24, Di - 24.97, En - 8.44, Ol - 11.87, Ap - 0.02.

Metaplagiogranite

A single sample of metaplagiogranite was analyzed (Table 6). The rock type occurs as fairly narrow dikes with sharp contacts within the metagabbro unit and consists of quartz, plagioclase and muscovite, with minor amounts of microcline, epidote and garnet. Although this metaplagiogranite contains more K_2O than the ideal described by Coleman and Peterman (1975), it is much less potassic and much more sodic and calcic than the typical granitic dikes of the southern Appalachian Piedmont. The low iron-magnesium, high silica and moderate alumina contents of the metaplagiogranite are consistent with oceanic plagiogranite described by Coleman and Peterman (1975). The original leucocratic intrusive rock probably formed by differentiation of the basaltic magma which also produced the pyroxenite and gabbro of the Hammett Grove suite.

Table 6. Analysis of metaplagiogranite.

Sample number	84F
SiO_2	73.70
Al_2O_3	15.47
TiO_2	0.04
Fe_2O_3 *	0.31
MnO	0.02
MgO	0.08
CaO	1.97
Na_2O	5.10
K_2O	0.92
P_2O_5	nd
Total	97.60

* Total iron as Fe_2O_3.

Normative mineral composition for metaplagiogranite (with iron assumed to be Fe_2O_3): Ilm - 0.04, Ab - 43.10, An - 9.78, Or - 5.44, En - 0.20, Crn - 2.49, Rt - 0.02, Qtz - 36.22, Hem - 0.31.

Metabasalt

Only one sample of the metabasalt was analyzed (Table 7). This sample was amphibolitic, consisted of hornblende and plagioclase and contained large, round plagioclase crystals interpreted as amygdules or feldspar phenocrysts. No pillow or sheeted dike structures have been recognized in the metabasalt of the suite, for the high grade of metamorphism and degree of deformation would have obliterated such features. Calculation of the normative mineralogy indicates a precursor for the metabasalt comprised of about 90 percent combined plagioclase

Table 7. Analysis of amphibolitic metabasalt.

Sample number	34
SiO_2	49.98
Al_2O_3	17.55
TiO_2	1.88
Fe_2O_3 *	11.90
MnO	0.20
MgO	5.61
CaO	11.54
Na_2O	2.37
K_2O	0.22
P_2O_5	0.14
Total	101.39

* Total iron as Fe_2O_3.

Normative mineral composition for metabasalt (assumed $FeO:Fe_2O_3$ = 3:1): Mag - 4.31, Ilm - 3.57, Ab - 20.03, An - 36.55, Or - 1.30, Di - 16.38, En - 16.18, Qtz - 1.79, Ap - 0.31.

and pyroxene with no normative olivine (Table 7). Lack of normative olivine commends this rock to a tholeiitic classification.

Major-oxide Plots

Figure 2 shows the normalized oxides plotted on an alkali-iron-magnesia (AFM) ternary diagram (with total iron in this and all subsequent plots reported as Fe_2O_3). The AFM diagram is often used to illustrate possible differentiation trends in assemblages of apparently related rocks (Wager and Deer, 1939). The altered ultramafite, except for one analysis of partially altered pyroxenite, plots distinctly separate from the metapyroxenites and metagabbro of the cumulate sequence. The altered ultramafite may represent the lowest part of an ultramafic cumulate zone, or possibly, may have been derived from the upper mantle. Although the metapyroxenites and metagabbro do not occupy distinctly separate fields, there is little real overlap of the metagabbro and metapyroxenite analyses; the metagabbro is characterized by moderate iron and alkali enrichment. Since only one sample of metabasalt was analyzed, it is difficult to draw definitive conclusions about the magmatic history of the extrusive portion of the suite, but it is obvious that the metabasalt is chemically distinctive with respect to the cumulate

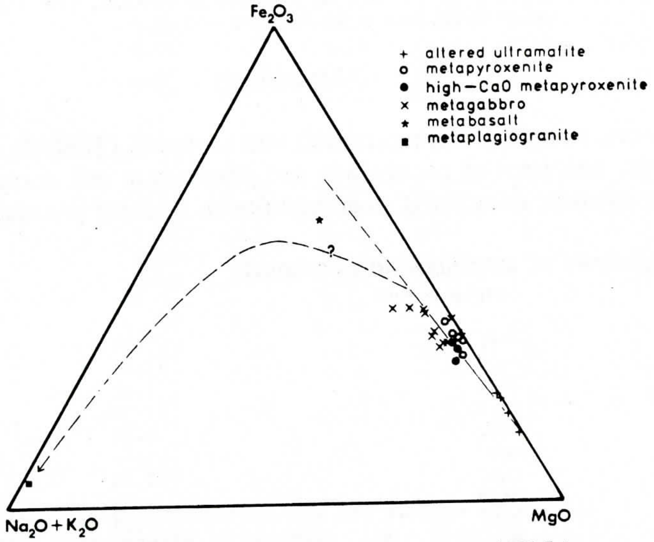


Figure 2. Analyses of Hammett Grove suite rocks on an AFM ternary diagram with possible differentiation trends.

sequence. The narrow compositional range of the metapyroxenites and metagabbro suggest that these cumulate rocks are comagmatic, a conclusion supported by intimate field relationships. A *possible*, bifurcating differentiation trend is shown in Figure 2, and it is recognized that paucity of metabasalt data *severely limits* this interpretation. The occurrence of metaplagiogranite dikes only within the metagabbro unit and the absence of any intermediate-composition rocks controls the illustrated interpretation of the differentiation trend extending from the gabbroic rocks near the iron-magnesium tie to the plagiogranite analysis near the alkali apex. In other words, since plagiogranite is restricted to gabbros of the cumulate sequence and because the gabbros and plagiogranite represent near-end member compositions on the AFM diagram, it is believed that the plagiogranite represents the final differentiation product of the magma that formed the ultramafic and mafic cumulate rocks. The distribution of compositions on the AFM diagram and the estimated differentiation trend are comparable to those of well-documented ophiolites (Bailey and Blake, 1974), especially the Papuan and Baltimore Complex ophiolites.

Figure 3 shows the normalized analysis plotted on a lime-alumina-magnesia (CAM) diagram. Coleman (1977) noted the interesting trend defined when analyses of metamorphic peridotites and mafic and ultramafic cumulates were plotted on the CAM ternary. He observed that the ultramafic cumulates, characterized by moderate Al_2O_3 and CaO contents, occupy a restricted field and do not overlap with metamorphic peridotites. Coleman also recognized that mafic cumulates overlap and partially follow the Skaergaard trend. Both of Coleman's observations are consistent with the CAM plot of Hammett Grove suite rocks. The general distribution of analyses within the CAM ternary diagram suggests that a magnesium-rich parent liquid differentiated to form the various and distinctive rock types of the suite, and that lime and alumina depletion characterized an early magmatic stage.

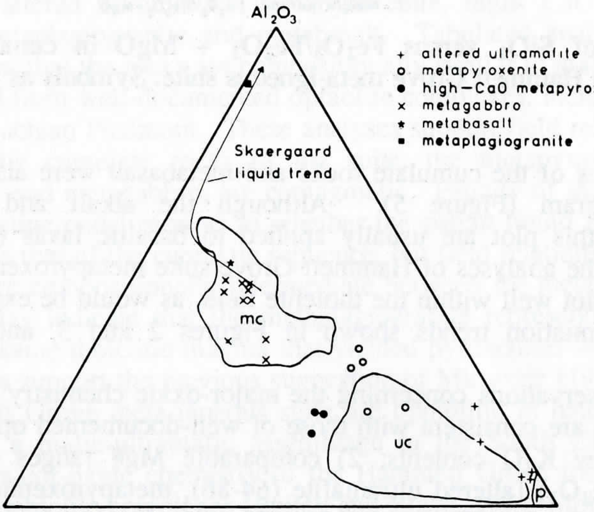


Figure 3. Analyses of Hammett Grove suite rocks on a CAM ternary diagram. Fields are from Coleman (1977). mc - mafic cumulate rocks, uc - ultramafic cumulate rocks, p - metamorphic peridotites.

Another oxide plot of some utility is the SiO_2 versus Fe_2O_3/Fe_2O_3 + MgO shown as Figure 4. Ultramafic cumulates, including both varieties of metapyroxenite, show little separation from the mafic cumulates (metagabbro). The narrow compositional range of the mafic and ultramafic cumulates of the suite further supports the interpretation that these rocks are comagmatic and are only moderately differentiated. The metabasalt analysis, although it probably does not adequately represent the extrusive part of the suite, plots well apart from the cumulate rocks, as in the AFM diagram. This chemical distinction may suggest that more complete fractionation of the parent magma occurred to form this basaltic rock, but alternatively, this separation may indicate that the basalts were not comagmatic, *sensu stricto*, with the cumulate rocks. A definitive conclusion will be possible only after more analyses, especially of the metabasalt, are completed.

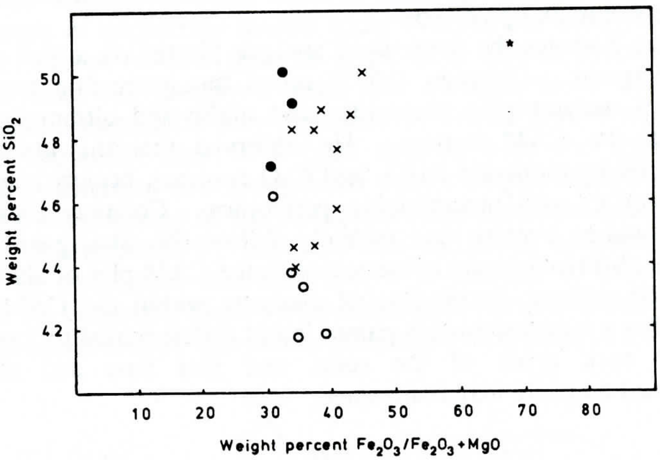


Figure 4. Plot of SiO_2 versus Fe_2O_3/Fe_2O_3 + MgO in cumulate rocks and metabasalt of the Hammett Grove meta-igneous suite. Symbols as in Figures 1 and 2.

The analyses of the cumulate rocks and metabasalt were also plotted on an alkali-silica diagram (Figure 5). Although the alkali and tholeiite field designations of this plot are usually applied to basaltic lavas (Macdonald and Katsura, 1964), the analyses of Hammett Grove suite metapyroxenite, metagabbro and metabasalt plot well within the tholeiite field, as would be expected for rocks with the differentiation trends shown in Figures 2 and 3, and is typical for ophiolites.

General observations concerning the major-oxide chemistry of the Hammett Grove rocks that are consistent with those of well-documented ophiolites include: 1) invariably low K_2O contents; 2) comparable Mg# ranges (Mg# = 100 x $MgO/MgO + Fe_2O_3$) [altered ultramafite (64-86), metapyroxenite (61-69), high-CaO metapyroxenite (66-70), metagabbro (55-66)]. Further, it should be pointed out that gabbroic rocks of the large Charlotte belt gabbro-metagabbro plutonic complexes to the east have relatively high soda and iron and low lime contents (McSween and others, 1984) as compared to the metagabbroic rocks of the Hammett Grove meta-igneous suite.

The trends of oxide increases and decreases in the Hammett Grove rocks are consistent with a model involving the simple crystal fractionation of original phenocrystic mineral phases. In general, with respect to decreasing MgO, iron, soda, potash, titania, lime and alumina increase from the metapyroxenites to the metagabbros (Figures 2 and 3). Silica varies somewhat systematically as is illustrated in Figure 4. These oxide trends are most simply explained by the fractional crystallization sequence olivine → clinopyroxene (± orthopyroxene) → calcic plagioclase.

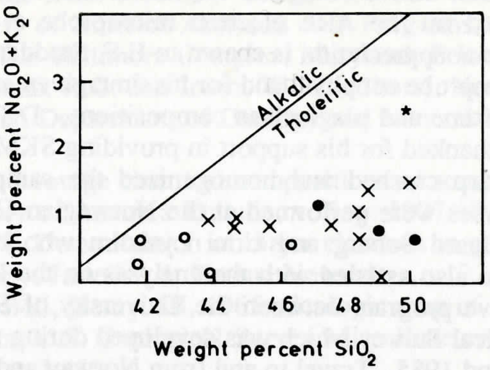


Figure 5. Alkali-silica plot of Hammett Grove suite cumulate rocks and metabasalt. Symbols as in Figures 1 and 2.

CONCLUSIONS

New major-oxide analyses demonstrate that the amphibolite-grade rocks of the Hammett Grove meta-igneous suite are divisible into six general groups interpreted as altered ultramafites, metapyroxenite, high- CaO metapyroxenite, metagabbro, metaplagiogranite and metabasalt. Tabulated analyses and various oxide plots show that the rocks are compositionally similar to mafic and ultramafic rocks described from well-documented ophiolite complexes, including those of the southern Appalachian Piedmont. These analyses support field relationships which indicate that the cumulate rocks of the suite, the metapyroxenite, high-CaO metapyroxenite and metagabbro, are comagmatic. Paucity of metabasalt analyses prevent a definitive conclusion as to whether the parent liquid for the metabasalt was the same as that of the cumulate sequence, but field relationships and the lack of rocks that are compositionally intermediate between the metaplagiogranite and metagabbros may indicate that the metaplagiogranite represents a differentiation product of the same tholeiitic magma that yielded pyroxenites and gabbros.

These data support the previous suggestion of Mittwede (1986) that the rocks of the Hammett Grove meta-igneous suite are ophiolitic in nature, the first such designation of a mafic-ultramafic assemblage in the South Carolina Piedmont. Although the major-oxide chemistry of the Hammett Grove rocks is clearly like that of known ophiolite complexes (Bailey and Blake, 1974; Morgan, 1977; Drake and Morgan, 1981; Kite and Stoddard, 1984), trace-element analyses and isotopic work are necessary to determine the tectonic-genetic environment in which the Hammett Grove suite formed. Presently, the writers believe that this mafic-ultramafic suite formed at the base of the Carolina terrane island-arc complex (Butler, 1987; Secor and others, 1986).

ACKNOWLEDGMENTS

The mapping and sampling aspects of this project were completed as part of a South Carolina Geological Survey (SCGS) mineral resources investigation.

Thanks to A.H. Maybin, III (SCGS) for his assistance in the early stages of the study of the Hammett Grove suite and for his continuing support. Microprobe work was performed on the ARL electron microprobe at North Carolina State University and special appreciation is shown to E.F. Stoddard for tolerating one of us (SKM) as a microprobe neophyte and for his time given and energy expended in obtaining clinopyroxene and plagioclase compositions. D.T. Secor (University of South Carolina) is thanked for his support in providing SKM with microprobe thin sections. Will Sharp crushed and homogenized the samples prior to analysis. The chemical analyses were performed at the Norwegian Geological Survey with the assistance of Baard Soberg and Unni Lysholm who carried out the sample dissolution and who also assisted with the analyses on the ICAP. This effort was part of a cooperative program between the University of South Carolina and the Norwegian Geological Survey which was developed during the visit of WES in the summers of 1984 and 1985. Travel to and from Norway and expenses while living in Trondheim were supported by a National Science Foundation travel grant (INT 8410912). We appreciate the insightful comments of D. Elthon (University of Houston and Lunar and Planetary Institute) and J.A. Karson (Duke University) who reviewed the manuscript.

REFERENCES

Bailey, E.H. and Blake, M.C., Jr., 1974, Major chemical characteristics of Mesozoic Coast Range ophiolite in California: U.S. Geological Survey Journal of Research, v. 2, p. 637-656.

Butler, J.R., 1987, Nature and significance of ultramafic bodies in the Piedmont of North Carolina and South Carolina: Geological Society of America Abstracts with Programs, v. 19, p. 77.

Coleman, R.G., 1977, Ophiolites, ancient oceanic lithosphere?: Springer-Verlag, New York, 229 p.

Coleman, R.G. and Peterman, Z.E., 1975, Oceanic plagiogranite: Journal of Geophysical Research, v. 80, p. 1099-1108.

Drake, A.A., Jr. and Morgan, B.A., 1981, The Piney Branch Complex-a metamorphosed fragment of the Central Appalachian ophiolite in northern Virginia: American Journal of Science, v. 281, p. 484-508.

Horton, J.W., Jr., 1981, Shear zone between the Inner Piedmont and Kings Mountain belts in the Carolinas. Geology, v. 9, p. 28-33.

Kite, L.E. and Stoddard, E.F., 1984, The Halifax County complex: Oceanic lithosphere in the eastern North Carolina Piedmont: Geological Society of America Bulletin, v. 95, p. 422-432.

Lieber, O.M., 1858, Report on the survey of South Carolina; being the second annual report: Columbia, South Carolina, 145 p.

Macdonald, G.A. and Katsura, Takeshi, 1964, Chemical composition of Hawaiian lavas: Journal of Petrology, v. 5, p. 82-133.

McSween, H.Y., Jr., Sando, T.W., Clark, S.R., Harden, J.T. and Strange, E.A., 1984, The gabbro-metagabbro association of the southern Appalachian Piedmont: American Journal of Science, v. 284, p. 437-461.

Mittwede, S.K., 1986, The Hammett Grove complex: Oceanic crust in the Piedmont of northwestern South Carolina: Geological Society of America Abstracts with Programs, v. 18, p. 256.

Mittwede, S.K., 1987, The Hammett Grove metaigneous suite and its tectonic implications for the Inner Piedmont- Kings Mountain belt boundary problem: Geological Society of America Abstracts with Programs, v. 19, p. 119.

Morgan, B.A., 1977, The Baltimore Complex, Maryland, Pennsylvania, and Virginia *in* Coleman, R.G. and Irwin, W.P., (editors), North American Ophiolites: Oregon Department of Geology and Mineral Industries Bulletin 95, p. 41-49.

Nockolds, S.R., 1954, Average chemical compositions of some igneous rocks: Geological Society of America Bulletin, v. 5, p. 1007-1032.

Ødegård, Magne, 1979, Determination of major elements in geologic materials by ICAP spectroscopy: Jarrell-Ash Plasma Newsletter, v. 2, p. 4-7.

Overstreet, W.C. and Bell, Henry, III, 1965, Geologic map of the crystalline rocks of South Carolina: U.S. Geological Survey Miscellaneous Geologic Investigations Map I-413.

Secor, D.T., Jr., Snoke, A.W. and Dallmeyer, R.D., 1986, Character of the Alleghanian orogeny in the southern Appalachians: Part III. Regional tectonic relations: Geological Society of America Bulletin, v. 97, p. 1345-1353.

Scotford, D.M. and Williams, J.R., 1983, Petrology and geochemistry of metamorphosed ultramafic bodies in a portion of the Blue Ridge of North Carolina and Virginia: American Mineralogist, v. 68, p. 78-94.

Sloan, Earle, 1908, Catalogue of the mineral localities of South Carolina: South Carolina Geological Survey, Bulletin 2, 505 p.

Wagener, H.D. and Price, Van, 1976, Geological reconnaissance maps of the Pacolet Mills and Wilkinsville quadrangles in Cherokee County: South Carolina Geological Survey Open-File Report 6.

Wager, L.R. and Deer, W.A., 1939, Geological investigations in East Greenland, pt. III, The petrology of the Skaergaard Intrusion: Kangerdlugssuaq, East Greenland. Medd. om Grønland, 105, p. 1-352.

Winkler, H.G.F., 1979, Petrogenesis of metamorphic rocks (fifth edition): Springer-Verlag, New York, 348 p.